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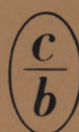
May, 1990

LITHOLOGY AND MINERAL RESOURCES

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A translation of *Litologiya i Poleznye Iskopaemye*

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TYPES OF MODERN HYDROTHERMAL AND HYDROTHERMAL-
SEDIMENTARY FORMATIONS IN THE ACTIVE ZONES
OF THE WORLD OCEAN

G. Yu. Butuzova

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In the article, the author distinguishes and characterizes the fundamental varieties of modern hydrothermal and hydrothermal-sedimentary formations on the floor of the World Ocean. A classification of such formations is proposed based on the following criteria: the ratio of hydrothermal material to exogenous matrix, the predominating types of chemical compounds in the ore matter, and the form and scale of its localization.

The most important scientific achievement of the recent decades in the field of geology was the discovery of hydrothermal and hydrothermal-sedimentary metalliferous deposits of iron, manganese, and polymetals on the bottom of the World Ocean.

Since the time of their discovery, investigators have obtained rich data base. The multidisciplinary study of this data base has allowed investigators not only to broaden but, in many respects, alter previously existing conceptions concerning oceanic metallogenesis.

The hydrothermal-sedimentary deposits on the bottoms of seas and oceans are generally characterized by an endogenous source of ore matter and deposition of this matter in a normal sedimentary manner; i.e., precipitation into a sediment, burial of the sediment, and redistribution over an area takes place in accordance with the geomorphology, hydrology, and physical-chemical environment of the sedimentary region. Deposits of such a nature are usually formed as a result of supplies of endogenous components entering into the composition of submarine exhalations and hot springs to the background material formed over the course of normal sediment formation. The sedimentary matrix, in turn, can be polygene and can consist of genetically diverse materials: terrigenous, biogenic, authigenic, and volcanogenic (pyroclastic, volcanoclastic, edaphogenic). Hydrothermal formations proper of the ocean floor are formed as a result of the same ore-forming process and differ from hydrothermal-sedimentary formations only with respect to the conditions associated with discharges of metalliferous solutions, discharges which take place in strata of bedrock and sediments not reaching the surface.

The broad range of conditions associated with the precipitation, burial, and distribution of ore components and the broad range of ratios of endogenous to normal-sedimentary matter suggest a significant variability and diversity of types and varieties of hydrothermal and hydrothermal-sedimentary accumulations in modern seas and oceans. This, in turn, demands that such accumulations be systematized and that classificatory principles be developed for them.

Several attempts to typify the hydrothermal-sedimentary formations of the active zones of the World Ocean have been undertaken. J. Edmond et al. distinguished three characteristic types of metalliferous deposits under the conditions of the mid-oceanic ridges: 1) deposits rich in Fe and Mn, 2) practically pure hydroxides, and 3) sulfide deposits rich in Fe but impoverished of Mn. It is assumed that all of these types are formed as a result of hydrothermal activity, while variations in composition are controlled by the degree of dilution of the high-temperature hot springs with normal seawater. The formation of sulfides, in the opinion of the authors just mentioned, occurs during minimal dilutions of primary solutions spilling directly onto the ocean floor. Crust-forming Mn hydroxides precipitate from cooling oxygen-containing solutions including only several percent primary fluoride in the compositions, while ferromanganoan sediments reflect intermediate stages in the process [32].

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At a conference on metalliferous oceanic spreading zones in 1978, D. Kronan proposed a subclassification of all the metalliferous deposits of the active spreading centers into three types: 1) deposits of sulfides associated with silicates and oxides (the ore muds of the Red Sea could serve as an example); 2) isolated, local deposits of silicates and oxides occurring primarily on mid-oceanic ridges; 3) widely distributed, predominantly oxide deposits of iron and manganese constituting the bulk of the metalliferous sediments of the mid-oceanic ridges [30].

The emergence of different types of metalliferous sediments is considered to be a result of a sequential, fractionated sedimentation of mineral phases from hydrothermal solutions. It is suggested in this case that sulfides precipitate during the earliest stages, followed by silicates, with metal oxides precipitating last.

On the basis of a classification of the hydrothermal deposits of the rift zones of the World Ocean, E. Bonatti proposed conditions associated with occurrences of deposits which could be used to distinguish groups of ore deposits: 1) stockwork-disseminated and massive sulfide deposits occurring within magmatic (predominantly ophiolite) complexes; 2) massive sulfide ores of the "black smoke" type, oxyhydroxide deposits of iron and manganese and silicate iron ores coinciding with the vent zones of hot-spring discharges onto the ocean floor; 3) deposits formed at some distance from the sources of the hot springs and subdivided into two groups (a: oxides and hydroxides of metals, silicates, and stratified sulfides of the Red-Sea type formed near the vents of hot springs, and b: oxides, hydroxides, and silicates displaced large distances from the discharge sites); 4) sheet deposits of hydroxides, silicates, and sulfides of various metals forming synchronously with the precipitation of terrigenous material within rift areas located far from dry land [26].

In this variant, as in the preceding variants, such important criteria as geochemical characteristics and actual mineral parageneses are not found, and it is important to note that the ratio of genetically diverse components, i.e., the metal-content associated with such components, is not taken into consideration. As a consequence, several varieties of deposits are now arbitrarily referred to as ores.

In general, the examples of formal typification discussed above are poorly suited for practical application when investigating oceanic metallogeny. The main reason for this is the absence of objective criteria to be used as a basis for carrying out a classification of the hydrothermal-sedimentary formations of the active zones of the World Ocean, which are exceptionally variegated in composition and distribution.

A more suitable and reasonable typification of deposits of this nature was proposed by E. G. Gurvich, Yu. A. Bogdanov, and A. P. Lisitsyn [7]. They distinguished and generally characterized the following classes of hydrothermal and hydrothermal-sedimentary accumulations within occurrences of oceanic crust and open ocean: 1) sulfide formations in the layer of the oceanic crust; 2) massive sulfide bodies at the ocean-floor surface; 3) ferromanganese hydrothermal crusts; 4) hydrothermal-sedimentary metalliferous formations (metalliferous sediments) of the seas and open ocean. Deposits of the deep-water basins of the Red Sea were treated separately as belonging to a water body detached from the ocean. The ancient metalliferous sediments of the ocean floor and the metalliferous formations of ancient spreading zones were also treated separately.

In contrast to the preceding classifications, a combination of such criteria as the conditions associated with localization, the compositions of the deposits and peculiarities of their geochemistries were also taken into consideration; however, unified classificational principles are absent for the most part. Different criteria underlie the identification of individual groups and there is no quantitative evaluation of the metal content within the deposits.

The absence of a unified typification for the broad spectrum of hydrothermal-sedimentary accumulations in modern seas and oceans created on the basis of clear classificational criteria especially complicates the comparative study of such accumulations and generates great terminological confusion when describing and characterizing this important group of mineral formations. For example, one and the same deposit associated with the deep-water basins of the Red Sea appears in the literature as ore-bearing, ore, and metalliferous sediments, without clarification of the geochemical and metallogenic meaning of these terms. At the same time, sediments containing no less than 10% Fe + Mn, calculated on the basis of carbonate-free material, have been referred to as metalliferous [12]. According to this definition,

TABLE 1. Average Contents of Al, Ti, Fe, and Mn (in carbonate-free material), % and Average Values of Geochemical Moduli in Different Types of Sediments and Rocks

Deposit and type of zone of the Red Sea (number of samples)	Al	Ti	Fe	Mn	Ti/Al	Fe + Mn	Al/Al + Fe + Mn	Published source
						Ti		
Basins of the rift zone of the Red Sea								[14, data of the author]
Ore-bearing sediments								
Atlantis-II (135)	0,97	0,07	35,7	1,41	0,07	530	0,02	
Tethys (35)	1,19	0,12	34,9	13,0	0,1	400	0,02	
Gypsum (25)	1,65	0,09	38,8	0,36	0,05	435	0,04	
Discovery (8)	0,87	0,08	55,8	0,41	0,09	700	0,015	
Albatross (9)	0,34	0,04	50,1	0,23	0,10	1300	0,016	
Erba (5)	0,5	0,05	50,0	0,2	0,10	1000	0,01	
Shagara (5)	0,5	0,03	35,4	18,6	0,06	1800	0,009	
Average (222)	0,86	0,07	42,9	4,9	0,08	683	0,015	
Metalliferous sediments								
Discovery (14)	5,36	0,5	20,3	2,4	0,09	45	0,19	
Albatross (10)	5,8	0,47	15,0	0,9	0,08	34	0,27	
Erba (17)	4,1	0,35	17,7	0,9	0,09	53	0,18	
Shagara (6)	6,2	0,56	11,6	4,4	0,09	29	0,27	
Nereus (19)	5,0	0,38	14,6	3,27	0,08	47	0,22	
Suakin (13)	5,3	0,43	16,5	3,5	0,08	46	0,21	
Average (79)	5,29	0,45	15,9	2,5	0,08	41	0,22	
Normal sediments (55)	6,5	0,6	7,8	0,7	0,09	14	0,43	
								[16, 28]
Eastern-Pacific Rise:								
Axis of ridge and fault zones (17)	1,98	0,18	24,0	8,25	0,09	180	0,06	
Flanks of ridge (111)	5,42	0,4	9,69	5,89	0,07	39	0,26	
Average	4,35	0,32	18,1	5,09	0,07	72	0,16	
Galapagos Rift Zone								
Fe-Mn deposits (13)	0,76	0,04	5,64	37,2	0,05	1070	0,01	[16, 38]
Silicate deposits-smectites (39)	0,95	0,06	20,5	0,5	0,06	350	0,04	
Mid-Atlantic Ridge FAMOUS area:								[37]
Fe-Mn deposits (2)	0,62	0,04	10,5	27,0	0,06	940	0,01	
Silicate deposits-smectites (7)	0,76	0,04	22,9	2,87	0,05	645	0,03	
TAG geothermal field:								[45]
Surficial metalliferous sediments	4,5	0,37	17,9	0,6	0,08	50	0,19	
Normal sediments	5,9	0,49	5,7	0,5	0,08	14	0,49	
Triple articulation of ridges in the Indian Ocean (219)	5,42	0,31	11,85	2,47	0,06	45	0,25	[15]

normal sediments containing hydrothermal material in the form of an admixture and accumulations almost completely consisting of ore components fall into a single category.

Taking into account that the hydrothermal and hydrothermal-sedimentary metallogeny of the World Ocean is one of the youngest subjects in geological science, development of the subject based on a complete classification of the mineral formations of this genetic type apparently is a matter for the future. However, an entire series of varieties, constituting in itself a schematic basis for a future, possibly more complete classification can now be identified and characterized on the basis of existing data.

To us, it seems most expedient to first classify all of the hydrothermal-sedimentary deposits of the active zones in the World Ocean according to their metal content, i.e.,

TABLE 1 (continued)

Fe-Mn sediments of the Gulf of Aden (6)	0,78	0,08	2,67	37,92	0,102	508	0,02	[29]
Cauldera of Santorin Island in the Aegean Sea (10)	0,40	0,02	37,4	0,09	0,05	1875	0,01	[1]
Banu-Wahu region (Indonesia)	2,1	0,15	32	5,5	0,07	250	0,05	[8]
Exhalational-sedimentary ores of Mn								[6]
Northern Kazakhstan, Ordovician (7)	1,52	0,13	5,2	35,4	0,085	312	0,03	
Eastern Bashkir, Devonian (11)	0,93	0,06	3,1	38,5	0,065	693	0,02	
Deep-water red clays	8,5	0,46	6,5	0,67	0,054	16	0,54	[47]
Pelagic clays of the Pacific Ocean	7,59	0,40	4,32	0,28	0,053	11	0,62	[16]
Clays of the platforms and geosynclines	8,96	0,49	4,73	0,05	0,054	10	0,65	[19]
Basalts of the mid-oceanic ridges								[9]
Eastern Pacific Rise	8,63	0,96	7,1	0,12	0,111	8	0,54	
Mid-Atlantic Ridge	8,68	0,84	7,4	0,12	0,097	9	0,53	
Ridges of the Indian Ocean	8,74	0,90	6,8	0,13	0,103	8	0,55	
Average	8,68	0,90	7,1	0,12	0,103	8	0,54	

according to the ratio of endogenous (hydrothermal, exhalational) ore matter to exogenous components, regardless of the actual features of the components involved.

A more detailed subclassification can be carried out on the basis of such criteria as the basic types of chemical compounds and the character of the localization of the sediments.

The most objective criteria allowing an evaluation of the degree of enrichment of sediments with endogenous material are geochemical moduli, among which the titanium (Fe + Mn):Ti [21] and aluminum Al: (Al + Fe + Mn) [28] moduli are widely used.

Underlying the modulus method of geochemical analysis are the fundamental properties of the chemical elements; these are reflected in migrational abilities and the formation of one or another of the mineral phases. Such elements-hydrolyzates as Al and Ti were recently shown to have low mobilities in hypergene, low-temperature processes by B. B. Polynov, N. M. Strakhov, A. I. Perel'man, N. A. Lisitsyn, and others. During an investigation of weathering processes under natural and laboratory conditions, it was thus shown that Al and Ti always occupy extreme positions associated with a number of the least migration-prone elements in the series of chemical mobilities of elements [11].

N. M. Strakhov emphasized the overall geochemical pattern characteristic of clays, sandy-silty rocks, and bauxites of various ages, which includes Al and Ti in a constant ratio [20].

An analysis of a large data base (my own and published data) allows me to assert that the same pattern is fully present in the modern metalliferous and ore formations associated with the active zones of the World Ocean, ancient hydrothermal-sedimentary ores, and also oceanic pelagic sediments.

This conclusion was made on the basis of the data on the distribution of Al, Ti, Fe, and Mn in modern sediments and in ancient ores presented in Table 1 and graphically demonstrated in Fig. 1.

A strong, direct correlation can be seen between Al and Ti (see Fig. 1). The correlation coefficient equals 0.977 at the 95% confidence interval. The figure shows an inverse correlation between Al and the sum of Fe and Mn, the chief components in terms of mass (the correlation coefficients equaled 0.864 at the 95% confidence interval). On the whole, the

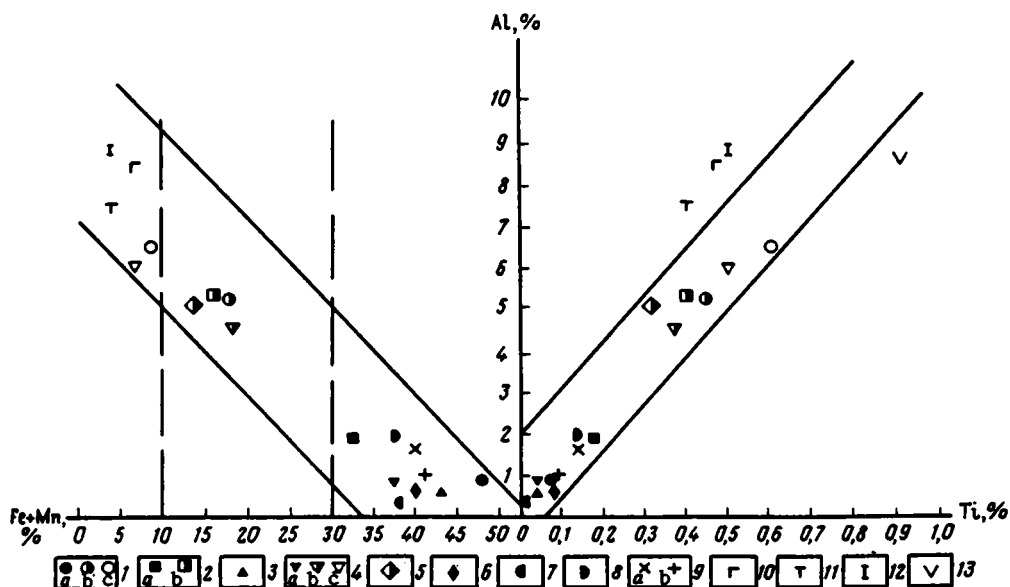


Fig. 1. Ratio between Al, Ti, and total ore components (Fe + Mn) in various types of sediments and rocks. 1) deposits of the rift zone of the Red Sea (a: ore-bearing, b: metalliferous); 3) ore-bearing deposits of the Galapagos rift zone; 4) deposits of the Mid-Atlantic Ridge (a: ore-bearing sediments of the FAMOUS area, b: metalliferous sediments of the TAG geothermal field, c: normal sediments of the TAG geothermal field); 5) metalliferous sediments of the Indian Ocean; 6) ore-bearing deposits of the Gulf of Aden; 7) ore-bearing sediments of the cauldron on Santorin Island; 8) ore-bearing deposits of the Banu-Wuhu region (Indonesia); 9) ancient hydrothermal-sedimentary manganese ores (a: Northern Kazakhstan, Ordovician, b: Eastern Bashkir, Devonian); 10) deep-water red clays; 11) pelagic muds of the Pacific Ocean; 12) muds of the platforms and geosynclines; 13) basalts of the mid-oceanic ridges.

low contents of Al and Ti are characteristic geochemical features of metalliferous and, especially, ore hydrothermal-sedimentary accumulations on the ocean floor; this was the reason such accumulations were initially referred to as "aluminum-poor ferromanganoan deposits" [28].

The paragenetic relationship between aluminum and titanium and also the regular decreases in the concentrations of these elements with increasing contents of the hydrothermal ore components iron and manganese which have been established for hydrothermal-sedimentary deposits fairly definitely indicate the presence of overwhelming portions of Al and Ti in the composition of lithogenic (terrigenous, volcanogenic) particles.

As follows from the data presented in Table 1, the ratio of titanium to aluminum in the clays of the oceans and continents averages 0.054; this ratio equals 0.103 in the basalts of the mid-oceanic ridges. The points in the figure corresponding to continental and oceanic clays occupy the extreme left position; the points corresponding to basalts of the mid-oceanic ridges are to the extreme right.

In the overwhelming majority of cases, the different types of metalliferous and ore formations occupy an intermediate position between clays and oceanic basalts with respect to Ti/Al values (see Table 1); this suggests a mixed composition for their lithogenic matrix, the chief components of which are terrigenous clays and products of basaltic volcanism.

The point concerning the very low migrational ability of Al and Ti in the hydrothermal process is confirmed by the results of direct analyses of thermal solutions of rift zones, which, unfortunately are not numerous thus far. The Al content within individual water samples is at the level of microconcentrations (units of mmole/kg); I did not find data concerning Ti in the literature.

The evidence of a predominantly lithogenic origin for the aluminum and titanium fully supports the use of geochemical moduli of these elements for uncovering the presence of endogenous ore matter in the sediments and for tentative estimates of the degree of concentration of these elements in the compositions of the metalliferous formations of the ocean floor.

It should be kept in mind that endogenous ore material can contain purely exogenous components sorbed from seawater by freshly deposited hydroxides of iron and manganese; however, the total quantity of such components is negligibly small in comparison with the quantities of endogenous iron and manganese. This allows us to neglect matter sorbed from the water when estimating the hydrothermal contribution to the total mass of sediment.

N. M. Strakhov, when discussing the distribution of elements over a broad spectrum of sedimentary and hydrothermal-sedimentary formations, arrived at the conclusion that the value of the titanium modulus for the sediments and rocks free of hydrothermal (exhalational) influence did not exceed 25, as a rule. According to the opinion of N. M. Strakhov, it is generally true that "values of the modulus significantly exceeding 25 ... should be considered reliable evidence for the inclusion of an exhalational component in a sediment" [22, p. 25].

Values of the aluminum modulus for normal sediments devoid of admixtures of ore matter are usually higher than 0.4, as our calculations show.

In Table 1, I present values of geochemical moduli for a broad spectrum of hydrothermal-sedimentary formations in comparison with normal sediments and rocks free of the influence of hot springs and also in comparison with separate types of ores. The table shows wide ranges of variation in the modulus values, from values insignificantly exceeding the terrigenous background to especially high (>1000) values for titanium and low (<0.01) values for aluminum; this reflects the large variability associated with ratios of normal sedimentary material and endogenous ore matter in metalliferous accumulations.

An analysis of the existing data indicates that, at values of the titanium modulus greater than 100 and values of the aluminum modulus are less than 0.1, the carbonate-free material usually contains more than 30% ore components (primarily Fe + Mn) and hydrothermal material constitutes no less than half of the sediment mass.

Taking these facts into consideration, it is proposed that all of the hydrothermal-sedimentary deposits can be subdivided into two large classes on the basis of values for the geochemical moduli and the content of ore matter: 1) ore-bearing, with a titanium modulus >100 and an aluminum modulus <0.1 , Fe + Mn $> 30\%$ (in the carbonate-free matter); 2) metalliferous, with a 25-100 range of values for the titanium modulus, a 0.4-0.1 range for the aluminum modulus, and a range of 10-30% for the content of Fe + Mn.

It is important to keep in mind that the term "ore-bearing" does not imply an estimate of economic value of a deposit, since only the ratio of ore to non-ore components of the sediment was being considered when distinguishing this class. The total mass or reserves of mineral crude is not taken into account; neither are the modern conditions of the deposit, nor the technology and cost involved in recovery. Therefore, potential ore deposits, insignificant ore shows, and ore mineralizations can also be included in the class of ore deposits. Note that D. Kronen also subdivides all metalliferous sediments into two groups on a purely qualitative basis: 1) normal, and 2) anomalous, i.e., local deposits sharply enriched in metals [10], which I refer to as ore-bearing.

I suggest that the most suitable account of the identified classes should be carried out on the basis of such criteria as the predominant type of chemical compounds and the character of the localization of the ore material.

An over-all scheme for the typification of the deposits discussed is presented in Table 2.

A brief characterization of the basic features associated with the material composition, morphology, and conditions of occurrence of each of the varieties distinguished is presented below.

ORE-BEARING DEPOSITS

Ob deposits can be subdivided on the basis of the predominant types of chemical compounds into sulfides, silicates, and oxides.

I. Sulfide Deposits. These deposits spatially and genetically are most closely connected with high-temperature hydrothermal activity. They are especially diverse with respect to mineralogy and geochemistry; they contain maximal amounts of such economically important metals as Cu, Zn, Pb, Ag, and several of their varieties are commercially promising.

Sulfide deposits can be divided into three subtypes on the basis of the conditions of occurrence, the morphology of the ore bodies, and the scale of distribution.

Massive Sulfide Bodies on the Ocean Floor. The widely used term in the literature, "massive sulfides," is a fairly arbitrary term, since it is not associated with clear morphological, physical, and structural-textural restrictions, and since it primarily designates cone-shaped, hilly structures located on the basaltic basement of the ocean floor or at the surface of the sedimentary stratum. The similar structures are usually complete with pipes or channels along which hydrothermal solutions either flowed or spill out of at present. The diameters of these cones vary from 1-2 to 20-30 m; the heights reach 50 m (usually 5-6 m) [34, 36].

One of the largest accumulations of sulfide ores is found within the Galapagos rift. This stratified deposit has a thickness of 35 m, a width of 200-200 m and extends for 1 km. It has been proposed that the sulfide body is formed as a result of growth of separate nearby cones dying off and the cones themselves crumbling [18, 40, etc.].

Hydrothermal structures in the active stage of the process are referred to as "smokers." Depending upon the temperature regime and discharge of the source, these structures can be subdivided into "black" and "white" smokers. Maximally high temperatures of thermal solutions ($>350^{\circ}\text{C}$), high (up to 24 m/sec) velocities of movement, and release of black, finely dispersed suspended material consisting primarily of pyrrhotite and iron shalerite are characteristic of black smokers. The temperatures of the hot springs of white smokers are significantly lower ($<350^{\circ}\text{C}$), and the discharge is weaker. The suspension released predominantly consists of an opalescent mass of amorphous silica with small admixtures of barite and pyrite [33].

The textural features, mineral compositions, and spatial interrelationships associated with the material composing the pipes and basal cones indicate the formation of the latter as a result of the destruction of ore-conveying channels and even partially as a result of the sedimentation of suspended particles. The "smoke" above the smokers rises several hundred meters. The bulk of the fine suspended material is dispersed during the water circulation and is oxidized [32, 34].

Samples collected from different areas of the cone-shaped structures are extremely non-uniform. Together with pipelike, honey-combed, and porous aggregates, interbeds and sheets of massive coarsely crystalline sulfides and samples with a clearly concentric zonality are encountered. Their compositions are especially complicated. The predominant minerals with respect to mass are sulfides of iron, copper, and zinc. The walls of the pipes are composed of sulfides and Ca and Ba sulfates (anhydrite barite). The hollows and pipes of dying smokers in the active areas are usually plugged with sulfide minerals.

Together with the sulfides, one finds within the composition of the hydrothermal cones sulfates (anhydrite, gypsum, barite, sulfates of Mg, Cu, Zn, etc.), silicates (predominantly amorphous silica, talc, ferruginous smectites, and, more rarely, zeolites), carbonates (calcite) and chlorides (atacamite).

Dead structures, where products of the oxidation of primary hydrothermal formations (Fe hydroxides, jarosite, elemental sulfur, etc.) are widely occurring, are characterized by the most complex mineral compositions. The walls of the pipe of active channels, as a rule, are dark gray overall and are characterized by a completely defined mineralogical, chemical, and textural zonality. In the interior, central parts of the channels, there are sulfides of Fe and Cu represented primarily by calcopyrite and cubanite. The exterior portion usually consists of fine-grained crystals of sphalerite, pyrite, marcasite, and pyrrhotite in association with anhydrite or barite. In a number of instances, the exterior zone is separated from the interior by an interlayer of bornite-calcopyrite composition. A similar characterization of the mineral composition of massive sulfides is presented in [24, 34-36, 42, and other reports).

Numerous chemical analyses of massive sulfides from different regions indicate an almost complete absence of exogenous terrigenous material within their compositions. Their components (Fe, Zn, Cu, Si, Mg, and S) constitute 100% of the composition as a rule.

TABLE 2. Scheme Showing Typification of Modern Hydrothermal and Hydrothermal-Sedimentary Formations of the Active Zones of the World Ocean

Group of deposits			Minerals		Chemical elements		Forms and scales of distribution	Typical examples of localization	Published sources
according to content of ore matter	according to predominant type of chemical compound	according to conditions of occurrence	principal	accessory	principal	accessory	Hilly structures up to 50 m in		
1	2	3	4	5	6	7	8	9	10
Ore-bearing (Fe + Mn)/Ti > 100, Al/(Al + Fe + Mn) < 0.1, Fe + Mn (in carbonate-free matter) > 30%	accessory	Massive	Pyrite, calcopyrite, marcasite, sphalerite, wurtzite, pyrrhotite, cubanite	Anhydrite, gypsum, barite, SiO ₂ (amorph), Fe-smectites, talc, saponite, Fe hydroxides. Mg, Cu, Zn sulfides, Cu oxychlorides etc.	Fe, Cu, Zn, S	Si, Mg, Ba, Pb, Cd, Ag, As, Sb, Au, Pt, Co, Ge, Se, P	Hilly structures up to 50 m in height, diameter to 30 m at sites of thermal-spring discharge	21° and 13° Juan de Fuca ridge, Galapagos frift, Gult of California (Guaymas basin)	[16, 18, 24, 28, 32, 34-36, 38, 42, and others]
		Lentiform-stratified	X-ray amorphous sulfides, Zn, Fe, Cu, sphalerite, pyrite, calcopyrite, pyrrhotite, Cu sulphosalts	Fe hydroxides (amorphous, goethite, hematite), Fe-smectidtes, SiO ₂ (amorph.), anhydrite, gypsum, manganosiderites	Fe, Zn, Cu, S	Si, Pb, Cd, Ag, Sb, Co	Interbeds and lenses within sedimentary stratum limited by the size of the basin-type	Red Sea, Salton Sea (?)	[2, 3, 14, 41, and others]
		Vein-disseminated in basalts	Calcopryrite, pyrite, and pyrrhotite	Fe-smectites quartz, chlorite, carbonates, zeolites, Fe-hydroxides. titanomagnetite (?)	Fe, Cu, S	Zn, Si, Ti	Veins from parts of a millimeter to 10 cm in diameter and micro-impregnation with in bedrock	Mid-Atlantic and Arabian-Indian ridges, Costa-Rican rift, isolated areas of East Pacific Rise	[17, 18, 27, 36, and others]
	Silicates	Lentiform-stratified	Nontronite, mix-layer Fe-smectites, celadonite	Sulfides, Fe hydroxides, SiO ₂ (amorph.), talc, saponite	Fe, Si	K, Mg, Al	Interbeds and lenses, consolidated coagulations and crusts in the sedimentary stratum	Red Sea, Gulfs of California and Aden	[2, 4, 29, 38, 48, and others]
		Areas within hydorthermally altered rock	Fe-smectites	Todorokite, bernessite, sedimentary carboantes	Fe, Si	K, Mg, Al	Irregular types of areas and nests	Lesser-Antilles island arc, Galapagos rift	Data of the authors and [48]

TABLE 2 (continued)

1	2	3	4	5	6	7	8	9	10
	Ferroman- ganoan oxides	Lentiform- stratified	X-ray amor- phous Fe and Mn hy- droxides, goethite, hematite, lepidocrocite, todorokite, bernessite, manganite		Fe, Mn	Si, Zn, Cu, Pb	Interbeds of various thick- ness and ex- tent within the sedimentary stratum	Red Sea, Galapagos rift, Mid- Atlantic ridge (TAG and FAMOUS fields), gulfs of California and Aden. active sub- marine vol- canoes (Santorin, Banu-Wuhu)	[1, 2, 8, 10, 14, 29, 31, 34, 37, 39, 44-46, 48, and others]
		Crusts on, surfaces of sediments and rocks	Todorokite, bernessite, goethite, hydrogoethite	Amorphous Fe and Mn hydroxides, Fe-smectides, SiO ₂ (amorphous)	Mn, Fe	Si, Ca, Mg	Hard crusts with thick- nesses of tens of centi- meters	Galapagos rift, Gulf of Aden, Mid-Atlantic ridge, East Pacific Rise, etc.	[5, 10, 18, 23, 25, 29]
		Areas within hydrothermally altered rock	Todorokite, goethite	Fe-smectites, Mn bernessite, Fe-hydroxides		Fe, Si, Ca, Mg	Irregular types of areas, nests, replacement of biogenic and volcano- genic parti- cles	Lesser- Antilles island arc, Galapagos rift	Data of the authors and [48]
Metalli- ferous (Fe + Mn)Ti, 25-100; Al/ (Al + Fe + Mn), 0.4- 0.1; Fe + Mn (in car- bonate-free matter), 10-30%	Ferroman- ganoan oxides	Covering, "basal" at the bottom of the sedimen- tary stratum	X-ray amor- phous Fe and Mn hy- droxides	Fe-smectites, Fe, Mn SiO ₂ (amorph.), goethite, todorokite, bernessite, vernadite, magnetite, gypsum, an- hydrite, barite, manganese- containing dendrites, Cu oxychlor- ides, etc.	Fe, Mn	Cu, Zn, Pb, V, Ni, Co, Mo. Sb, U, Zr, P, Ba REE	Wide area (up to several mil- lions of square kilo- meters) of mid-oceanic ridges, iso- lated basins of rift valleys	Southeastern Pacific Ocean, Mid- Atlantic and Mid- Indianridges, basins of the Red Sea, etc.	[5, 7, 10, 12-16, 21, 23, 28, 31, 32, 44, 45 others]

Massive sulfides are now known to be present at many points along the mid-oceanic ridges, predominantly in the Pacific Ocean within zones with average and high spreading rates (21 and 13° N, 20-21.3° S, Juan de Fuca ridge, Guaymas basin in the Gulf of California, the Galapagos rift).

Lentiform-Stratified Deposits of Sulfides. Layered sulfide deposits occurring within the sedimentary stratum or at its surface and formed as a result of the discharges of hot springs onto the floors of geomorphologically pronounced deep-water basins are known to exist only within the rift zone of the Red Sea according to present data.

This is connected with a specific circumstance associated with ore formation, to be precise, it is connected with the presence of oxygen-free, highly mineralized waters (brines) at the discharge sites of hot springs. Dense brines filling geomorphologically pronounced depressions of the sea floor impede the penetration of surface waters to the bottom and thus create conditions favorable for the preservation of hydrothermal sulfides within the sediments by protecting such sulfides from oxidation.

A characteristic feature of the sulfide mineralization in the Red Sea is the irregularity associated with the distribution of sulfides, both vertically within the sediments and over the area of the basin. Nevertheless, horizons sharply enriched in sulfides can be distinguished in the section of the sedimentary stratum. Periods of maximum intensity of hydrothermal activity and the ore-forming process are responsible for the formation of these horizons. The sulfides within such horizons do not form individual sheets maintained along course, but are irregularly disseminated within a mass of siliceous-ferruginous or anhydrite deposits, forming piles, crusts, lenses, and finely dispersed microimpregnations. The interbeds and lenses enriched in sulfides are distinguished by a sooty-black color and are characterized by very low (from +50 to -90 mV) Eh values for ore-bearing deposits. In the ore-bearing sediments of the Red Sea, the most widely occurring sulfides of iron and zinc are represented by x-ray amorphous phases, pyrite, and sphalerite. The highest concentrations of sulfides, the highest mineralogical diversities of sulfides, and best crystallizations of sulfides are characteristic of areas located in direct proximity to the discharge sites of hot springs, where, along with pyrite, sphalerite, and x-ray amorphous phases, pyrrhotite and the sulfide compounds of copper (calcopyrite, copper sulphosalts) are also widely occurring [2]. The sulfide interbeds are distinguished by maximum contents of such elements as Cu (up to 2.5%), Zn (up to 10%), Pb (up to 0.4%), Ag (up to 0.02%), and Cd (up to 0.07%) found in the ore strata.

In the sulfide-containing horizons of the Atlantis-II basin, a fully defined geochemical zonality can be traced. Thus, a higher copper content in the sulfides marks areas associated with discharges of thermal solutions. Copper mineralization occurs in areas of limited size; zinc sulfides are distributed in a significantly broad fashion over almost the entire basin; lead sulfides are generally uncharacteristic of the ore sediments. High contents of lead are associated with involvement of the element in the compositions of iron, zinc, and copper sulfides.

The average value of the titanium modulus for the sulfide deposits of the Atlantis II basin amounts to 545; the average value for aluminum amounts to 0.07. The relatively low modulus values for titanium and the relatively high values for aluminum are connected with the fact that the composition of the ore matter includes, along with iron, a large quantity of such elements as zinc and copper, which are not taken into consideration during calculations of the geochemical moduli. The concentration of titanium in the sulfide interbeds does not exceed hundredths of a percent.

There is a complex combination of mineral phases associated with the sulfides in the Red-Sea sediments, including oxides and hydroxides of iron (amorphous, goethite, hematite), amorphous silica, ferruginous smectites, sulfates (anhydrite, gypsum), and carbonates (manganosiderites). The presence of small quantities of manganese are associated with the latter in the sulfide deposits of the Red Sea.

We cannot rule out the possibility that sulfides occurring in the Pliocene-Pleistocene deposits of the Salton-Sea geothermal system may be similar to the Red-Sea deposits with respect to composition, character of deposition, and genesis. However, the existing data in the literature [41] is still insufficient to confirm or deny this suggestion.

Vein-Disseminated Sulfide Mineralization.* This type of ore mineralization, as in the case of the massive sulfides on the floor of the World Ocean, is restricted to areas in which hydrothermal systems are present; the mineralization is directly associated with the hydrothermal systems and occurs in the stratum of oceanic crust, predominantly within the second (basaltic), more rarely at the top of the third layer. Vein disseminated sulfides are essentially hydrothermal rather than hydrothermal-sedimentary formations, since they form during the discharge of metalliferous hot springs in a stratum of bedrock. Signs of secondary vein-disseminated mineralization in oceanic basalts, found both along material from deep-water drilling and when dredging rocks, have been identified in a whole series of areas of rift zones, predominantly within regions with average and low spreading rates (the Mid-Atlantic and Arabian Indian ridges, individual areas of Eastern Turkistan, the Costa-Rican rift, etc.).

The sulfides in oceanic basalts usually form veins of fantastic shapes and varying lengths. The veins are no more than 5-10 cm in thickness and are usually significantly thinner (several millimeters). Alongside the vein mineralization filling the fissures in the rocks, sulfides are found in the form of finely disseminated, local microimpregnations, usually in the interstices between veins. In general, the scale of development of secondary hydrothermal sulfides is insignificant within the oceanic crust. These sulfides differ significantly from the massive sulfides of the ocean floor and from the essentially magmatic sulfides of magmatic bed rock, both in terms of composition and mineral paragenesis.

The disseminated and vein ore minerals are primarily pyrite, calcopyrite and pyrrhotite, i.e., iron and copper sulfides which occur in associated with secondary silicates (smectites, chlorite, quartz), more rarely zeolites, carbonates, and magnetite [27, 36, etc.].

The character of the ore mineralization in the bed rocks on the slopes of the Arabian-Indian ridge was described in detail by T. V. Rozanov and G. N. Baturin [17]. It was shown that the primary minerals, calcopyrite, and pyrite, form a vein dissemination in the quartz-chlorite vein mass. In this case, the structural peculiarities of the calcopyrite are characteristic of its high-temperature varieties.

In the basalts of the equatorial portion of the Mid-Atlantic Ridge within the zone of the ridge's intersection with the Romanche fault, the secondary sulfide minerals in the form of veins and isolated aggregates are also represented by calcopyrite, pyrite, and pyrrhotite [17].

The general geochemical specification of hydrothermal vein-disseminated sulfides consists in the widespread development of iron-copper mineralization and the sparse presence of zinc sulfides; this distinguishes hydrothermal vein-disseminated sulfides from passive occurrences and stratified sulfide deposits.

II. Silicate Deposits. Authigenic layered silicates, among which the group of ferruginous smectites predominate, are one of the characteristic mineral varieties of ore-bearing sediments. Their presence has been identified in the majority of the hydrothermally active regions of the World Ocean (Eastern Turkistan, the Red Sea, the Mid-Atlantic Ridge, the Gulf of Arabia, the Lesser Antilles island arc, etc.). Ferruginous smectites are most widely occurring in the deposits of the Galapagos rift.

Silicate deposits can be divided into two subtypes on the basis the character of the occurrence and the conditions of formation: hydrothermal-sedimentary stratiform deposits forming lentiform interbeds, and hydrothermal deposits proper infilling areas in hydrothermally altered rocks of various compositions with fantastic forms. The former are created as a result of the precipitation of amorphous or slightly crystallized siliceous-ferruginous phases from the bottom waters followed by diagenetic transformation of the phases precipitated; the latter are created during the discharges of hot springs prior to their exit onto the ocean floor, i.e., within sediments and rocks.

Lentiform-Stratified Silicate Deposits. Hydrothermal-sedimentary ferruginous smectites usually occur in a stratum of sediments in the form of lenses and lentiform interbeds distinguishable within the section by green coloration of diverse tint; these lenses and interbeds are compacted and form weakly lithified lumps, crusts, and rinds.

*In individual reports, formations of this nature may be referred to as mineralization of the "stock-work" type [18].

Tetrasilicic nontronites, i.e., highly ferruginous layered silicates with weak replacement of S by Al in the tetrahedral layers, are the most common mineral phases of smectites.

A detailed investigation of the silicate component of hydrothermal-sedimentary genesis was carried out for the ore-bearing sediments of the Atlantis II basin (the Red Sea). As a result, it was shown that the structurally disordered nontronites of the upper horizons are transformed, during gradual fixing of calcium from the muddy waters, into celadonite through a series of mixed-layer formations over the course of the aging and subsequent recrystallization of the siliceous-ferruginous gel precipitating from the solution. The mechanism and dynamics of this process under natural conditions were worked out in reports [4 and others].

In a still more detailed investigation of a group of ferruginous smectites in the region of the Galapagos rift [48], a clear tendency toward celadonization of the nontronites with increasing depth of the sedimentary stratum was also noted. It can be suggested that the mechanism of hydrothermal-sedimentary smectite-formation worked out for the examples of the Red Sea and Galapagos rift zone are common to all of the active regions of the World Ocean.

In areas situated in direct proximity to hydrothermal sources, a still more diverse complex of silicate minerals can occur. Thus, a complex of magnesium-containing silicates (talc, saponite, and chlorite) were found in the Atlantis II basin at discharge sites together with the usual layered silicates of the nontronite-celadonite series [12]. The magnesium silicates are prevalent in the deposits of the Gulf of California (Guaymas basin). Talc is a typical mineral admixture in the mass of the sulfide bodies on the ocean floor; large-scale occurrences of hydrothermal-sedimentary talc were found within individual regions of the Gulf of California near sources [39]. If a narrow spatial and genetic link with sulfide compounds is characteristic of silicate minerals enriched in magnesium, then ferruginous smectites are encountered in association with a still broader spectrum of hydrothermal-sedimentary formations of both oxide and sulfide specialization.

From a chemical point of view, ore-bearing deposits of a silicate composition are fairly homogeneous and are characterized by high contents of iron and silica in the presence of low concentrations of aluminum, titanium, manganese, and microelements, including such typical hydrothermal microelements as copper, zinc, and lead.

The values of the titanium modulus in silicate (smectite) interbeds and lenses are, as a rule, significantly higher than 100 (350 for the Galapagos region, 645 for the FAMOUS area); this allows us to confidently classify the interbeds as a type of ore-bearing sediment.

Silicates in Hydrothermally Altered Rocks. This variety of hydrothermal formations is still very poorly studied. We brought up indigenous volcanogenic-sedimentary rocks locally and intensively reworked by thermal solutions while dredging in the region of the Lesser Antilles island arc (the Zombi Fracture Zone, near the island of Guadeloupe).

Layered silicates forming areas and nests of fantastic, green-colored features are one of the principal hydrothermal neogeneses in the rocks. They are practically monomineralic and consist of microglobular segregations and solid masses of tetrasilicic nontronite.

The presence of newly formed smectites was also noted in the section of the sedimentary stratum of the Galapagos rift [48]. In this case, it was shown that ferrismectites and celadonite are formed in the pore spaces during the discharge of solutions and that Fe-Al smectites are formed during the interaction of the hot springs with the siliceous-calcareous matrix.

Hydrothermal smectites, as in the case of hydrothermal-sedimentary smectites, are characterized by extremely low contents of trace elements (Al, Ti, and microelements, including rare-earth elements).

Taking into account the mineral compositions and geochemical features of both varieties of smectite neogeneses and also the possibility of their coexistence within the same sedimentary section, the boundaries between the two varieties are not always clearly distinguishable. The most characteristic feature of smectites formed during the discharge of hot springs onto the ocean floor consists in the clearly expressed tendency toward variation in the composition, structure and degree of crystallization of the smectites over the course of the diagenesis from top to bottom along the section of the sedimentary stratum (celadonitization). The hydrothermal smectites in altered sediments and rocks are irregularly distributed and exist in complex relationships with other mineral neogeneses. Further investigations will probably allow us to clarify the mineralogical-geochemical and morphological differences between the groups of smectites under consideration.

III. Ferromanganoan Oxide Deposits. Hydrothermal-sedimentary and hydrothermal formations consisting predominantly of oxides and hydroxides of iron and manganese compounds are the most prevalent type of ore-bearing sediments. They occur in practically all of the active regions of the World Ocean.

On the basis of a whole series of criteria (shape of the deposit, mineralogical-geochemical features, scales of distribution and conditions of occurrence) these deposits can be subdivided into three types: 1) lentiform-stratified unconsolidated sediments in the stratum of sediments and at the sediment surface; 2) hard surficial crusts present on a different substrate of ocean floor; 2) formations occurring within hydrothermally altered rocks.

Lentiform-Stratified Deposits. Ore-bearing deposits primarily consisting of iron and manganese hydroxides predominantly occur in the trough regions of the rift valleys (the Red Sea, the gulfs of California and Aden), within the axial portions of mid-oceanic ridges, and also in regions of submarine volcanic activity of certain island-arc systems (the cauldrea of Santorin Island in the Aegean Sea, the submarine volcano of Banu-Wuhu in Indonesia, etc.).

The most favorable environments for the accumulation of ferromanganoan deposits are geomorphologically pronounced bottom depressions, which are natural traps for ore matter. As a rule, such depressions are located at sites of hot-spring discharges or in direct proximity to them. The ore accumulations usually contain an admixture of biogenic and terrigenous components. The values of the geochemical moduli vary of a wide range depending upon the quantity of such components. For example, the average value of the titanium modulus equals 180 for the sediments associated with the axial portion of the Eastern Pacific Rise, ~700 for the oxide deposits of the Atlantis II basin (the Red Sea), and >1000 for the ore accumulations in the harbors of the Santorin cauldrea.

Within the stratum of sediments, ore-bearing deposits in the form of interbeds and lenses of various thicknesses irregularly interfinger with normal muds, which can either include some quantity of hydrothermal material or are completely devoid of such material.

The mineral compositions of the ferromanganoan accumulations are diverse. Together with the widely occurring x-ray amorphous phases, most usually the following can be found among the ferruginous minerals: goethite, hydrogoethite, lepidocrocite, and hematite; more rarely magnetite and ferrihydrite can be found. Todorokite usually predominates among the manganese phases; bernessite, buzerite, vernadite, manganite, and rarer minerals such as asbolan and mixed ferromanganoan phases (goethite-groutite) also occur. Further investigations will probably allow us to refine and extend the list of ferromanganoan minerals in ore-bearing sediments.

The ratio of Fe to Mn in this type of deposit also varies widely, from almost pure accumulations of Fe or manganese to Fe-Mn sediments mixed in various proportions. The ratio of the principal ore components of iron and manganese is determined by the degree of fractionation of these components; this, in turn, depends upon a whole series of factors (the source flow, the flow regime associated with the ore process, physicochemical and geomorphological conditions associated with the deposition of the material, etc.).

It is known that separation of Fe and Mn during geochemical processes is a widely occurring phenomenon in nature and one clearly manifested during hydrothermal-sedimentary ore genesis. This is connected with the different migrational abilities of the elements resulting from their differing oxidation potentials. As a rule, iron is oxidized and precipitates in direct proximity to the discharges of the hot-springs, making pure iron-ore deposits fairly local in character. Under special conditions, manganese-ore interbeds can also be deposited near source vents (the Red Sea); the reason for this phenomenon is explained in [3]. Manganese, which is usually more mobile, and capable of migrating further in solution, settles out at significant distances from sources. The formation of mixed ferromanganoan deposits is due primarily to the mechanical transport of particles and also to processes involving coprecipitation and sorption. In the presence of other, equivalent conditions, the Fe/Mn values in the sediments generally decrease in a regular fashion with increasing distance from the discharge sites of the hot springs.

A common geochemical features of ferromanganoan ore-bearing deposits consists in their enrichment with the same group of elements which form sulfide deposits, i.e., elements entering into the composition of thermal solutions, principally copper, zinc, and lead, with other microelements in small concentrations.

TABLE 3. Primary Differences between Hydrothermal and Hydrogenic Crusts

Type of crust	Rate of growth, mm per 10 ⁶ yrs [18, 29, 44]	Content of microelements [23, 25, 29, 44]	Mn/Fe [25]	Fe/Mn [43]	Co/Zn [46]	U/Th [43]	Ce/La [5]	SiO ₂ (absolan) [46]	Primary minerals of the ore matter [10, 23, 29, and others]
Hydrothermal	100-1000	Low	>5 (to 50 and higher)	12-237 in ferrous, 0.002-0.14 in manganese ~1	<0.15 >1		<2	Much	Todorokite, bornessite, goethite, and hydrogoethite
Hydrogenic	1-10	High ~1			~2.5 <1		>2		

Ferromanganoan Crusts. It is known that hard crusts and incrustations consisting of oxide and hydroxide species of iron and manganese are widely occurring at the surface of the ocean floor. Two genetic varieties can be distinguished among them: hydrothermal crusts formed primarily from material arriving at the bottom within the compositions of exhalations and hot springs, and hydrogenic crusts growing as a result of precipitation of components from seawater. The question concerning the genesis of these natural formations is closely connected with the general question of the source and also the balance of forms of iron and manganese in the ocean and is still far from being definitely answered.

Nevertheless, important differences between crusts of different origins have been established on the basis of a whole series of indices (Table 3). These differences are determined primarily by the intensity of the delivery of ore components and the rate of their accumulation.

From the data presented in Table 3, it follows that the rates of growth associated with the hydrothermal crusts are approximately two orders of magnitude higher than the rates associated with hydrogenic crusts. The ferromanganoan crusts with the highest growth rates coincide, as a rule, with hydrothermally active areas of the ocean floor (the Eastern Pacific Rise, the Mid-Atlantic Ridge, the Galapagos rift, the Gulf of Aden, etc.). In contrast to hydrogenic crusts and films occurring over larger areas of the ocean floor, the hydrothermal formations are generally characterized by lower contents of the majority of microelements, including total rare-earth elements [23, 25, 29, 44, etc.]. This overall geochemical specification is a natural consequence of the high rates of accumulation of ore matter, the limited time of contact of the sorptionally active particles of Fe and Mn hydroxides with seawater, and the extraction of microelements from the seawater.

In comparison with hydrogenic crusts, hydrothermal crusts are more impoverished in microelements by a factor of 200-300 [25] and are more impoverished in rare-earth elements by one or two orders of magnitude [46]. As I. I. Volkov correctly pointed out, however, such an index as the level of concentration of microelements, as in the case of the quantity of rare-earth elements, cannot always serve as a reliable criterion of origin, depending, as they do, upon the ratio of the oxide species of Fe and manganese within the crusts. In his opinion, the composition of rare-earth elements, which is similar to the composition of the deep bottom waters and which is characterized by low values for Ce/La and La/Sb, is more informative for genetic interpretations [5].

The ratio of iron to manganese within the oxides of crusts and incrustations serves as an important indicator of the origin of these features (Table 3). Whereas a mixed ferromanganoan composition is characteristic of hydrogenic crusts, one or another element

usually predominates sharply in hydrothermal crusts. This predominance is determined by the fractionation of iron and manganese during the hydrothermal ore-forming process.

The mineralogy of ferromanganoan crusts and incrustations of the ocean floor is still not sufficiently studied. However, the existing data allows us to thoroughly outline specific differences with respect to the mineral compositions of genetically different formations.

Todorokite and bernessite usually predominate among the manganoan minerals of the hydrothermal crusts. Goethite and hydrogoethite predominate among the ferruginous minerals. The hydrogenic crusts are characterized by a more complex mineral composition; an extensive combination of ferromanganoan mineral phases, Fe-Mn vernadite, bernessite, hydrogoethite, etc., can be identified in such crusts.

Taking into account that the formation of crusts and incrustations takes place as a result of the precipitation of ore components from seawater, elements arriving from different sources can apparently exist within the same formation. Nevertheless, the entire combination of criteria noted above allows us to isolate hydrothermal crusts within a separate genetic variety.

Oxide Formations within Hydrothermally Altered Rocks. Oxide phases, developing in a manner similar to hydrothermal silicates within the stratum of bed rock as a result of discharges of thermal solutions, have been quite insufficiently investigated. We will include only data on the hydrothermally altered rocks of volcanically active regions of the Lesser Antilles island arc. The principal ore components within the cgc-sedimentary matrix are Mn hydroxides in the form of collomorphs, reniform and dendritic separations, and also areas of replacement of biogenic and volcanogenic particles. The Mn minerals occur in complex, unpredictable ratios with smectitic material.

The ore mass consists predominantly of well crystallized todorkite; they contain practically no admixture of chemical elements with the exception of Mg, Ca, and K. Against a background of ore matter consisting of hydroxide species of Mn, one encounters sparse areas enriched in iron (Fe to 20%). Unfortunately the mineral species of the oxide phases of the iron have still not been identified.

METALLIFEROUS DEPOSITS

In contrast to the ore-bearing deposits, the metalliferous sediments are more homogeneous, both with respect to composition and character of occurrence. Moreover, they have been studied in less detail; therefore, there is no basis at present for carrying out a more detailed classification.

The metalliferous sediments include a broad spectrum of normal muds consisting of terrigenous, biogenic, authigenic, and volcanogenic components which contain endogenous (ore) matter within their compositions.

The variations in the ratios of hydrothermal components to "endogenous matrix" are reflected in the magnitudes of the geochemical moduli (titanium 25-100; aluminum 0.4-0.1).

The general pattern associated with the localization of the metalliferous sediments consists in the coincidence of such sediments with tectonically active regions of the ocean floor, primarily coincidence with the axial and slope areas of the mid-oceanic ridges and the regions adjacent to them.

These sediments are most widely distributed in the southeastern portion of the Pacific Ocean, where they occupy an area of the order of 10 million km² [13]. Metalliferous sediments also occur in the Indian Ocean and the West-Indies ridges over an area of approximately 2 million km² [15], within individual areas of the Mid-Atlantic Ridge (the TAG geothermal field, the FAMOUS area), and within several basins of the Red Sea.

The large areas of occurrence of metalliferous sediments are due to the dissemination of hydrothermal components over the bottom surface by bottom currents. Maximum accumulations of ore components in the low regions of the sea bottom (in individual basin-traps and in rift valleys) are associated with this process.

In [23], the allochthonous nature of the ore matter in the metalliferous sediments is confirmed and the possibility of the occurrence of such sediments at fairly large distances from the sites of discharges from the hot springs is noted.

The hydrothermal component of the metalliferous sediments is itself primarily a mass of fine particles of iron and manganese hydroxides (tenths and thousandths of a millimeter). The major portion of the hydrothermal component is x-ray amorphous. The presence of ferruginous smectites, berthierite, vernadite, todorokite, goethite, and hydrogoethite has been noted among the crystallized phases. Such minerals as magnetite, cristobalite, gypsum, anhydrite, manganosiderites, barite, native metals, etc. have been noted in small quantities [13, 23, 31, and others]. In the opinion of L. E. Shterenberg, all of these minerals, as in the case of the ground mass of ore matter, was transported by aqueous currents from regions of hot-spring discharge. We found significant quantities of copper oxychlorite (atacamite) in the metalliferous sediments of the TAG geothermal field (Atlantic Ocean).

In comparison with the normal pelagic clays of the ocean, the metalliferous sediments are enriched in a whole series of microelements (Cu, Zn, Pb, Ni, Co, V, Mo, Sb, rare-earth elements, etc.) in addition to Fe and Mn and are poor in such lithophilic elements as Al and Ti. They are also characterized by a broader spectrum of microelements and a high level of microelement concentrations in comparison with hydrothermal-sedimentary ore-bearing deposits. This geochemical characteristic of metalliferous sediments is due to the prolonged contact of freshly formed hydrothermal particles of Fe and Mn hydroxides with seawater during the course of the transport of these particles from the source vents to the burial sites and processes of coprecipitation and sorption of microelements from bottom waters. In other words, the primary portion of microelements in metalliferous sediments has a purely exogenous (hydrogenic) source; both the concentrations of elements and their geochemical spectrum and the modulus calculations indicate this [16]. We did not exclude the possibility that further investigations of the ore matter during a first look into homogeneous metalliferous sediments might permit the discovery of certain variations in the mineral and chemical compositions of such sediments, might permit them to be subclassified into a number of varieties, and might explain the genetic origin of each. Apparently, the so-called basal ferromanganese deposits, which are genetically related to the type of deposit described and similar in composition, occur at the base of the sedimentary mantle of the oceans, predominantly on the flanks of the mid-oceanic ridges.

In conclusion, it should be emphasized that the proposed scheme for the typification of modern hydrothermal and hydrothermal-sedimentary oceanic formations probably does not preserve all of the diversity associated with such formations; moreover, the boundaries between certain subdivisions, as was pointed out, are still fairly tentative in nature.

Nevertheless, it is impossible not to conclude that each mineral variety reflects important genetic features of a single ore-forming process. Such features include the intensity and temperature regime, the composition and conditions of discharge of thermal solutions, and physical-chemical, geomorphological, and hydrodynamic circumstances associated with the precipitation and accumulation of the ore matter. We therefore propose that a systematization of the compositionally varied deposits of this genetic type will facilitate an understanding of oceanic ore genesis in general and also could be useful for estimating the commercial promise of each variety and for comparing such varieties with ore accumulations of the geological past.

LITERATURE CITED

1. G. Yu. Butuzova, "Modern volcanogenic-sedimentary iron-ore process in the cauldern of the Santorin volcano (Aegean Sea) and its influence upon the geochemistry of the sediments," Tr. Geol. Inst. Nauk Akad. Nauk SSSR, No. 194 (1969).
2. G. Yu. Butuzova, "Mineralogy and certain aspects associated with the genesis of metalliferous sediments of the Red Sea, Part I," Litol. Polezn. Iskop., No. 2, 3-22 (1984).
3. G. Yu. Butuzova, "Mineralogy and certain aspects associated with the genesis of metalliferous sediments of the Red Sea, Part II," Litol. Polezn. Iskop., No. 4, 11-32 (1984).
4. G. Yu. Butuzova, V. A. Drits, N. A. Lisitsyna, et al., "Dynamics associated with the formation of clay minerals in the ore-bearing sediments of the Atlantis-II basin (Red Sea)," Litol. Polezn. Iskop., No. 1, 30-42 (1979).
5. I. I. Volkov and A. V. Dubinin, "Rare-earth elements in the hydrothermal accumulations of iron and manganese in the ocean," Litol. Polezn. Iskop., No. 6, 40-56 (1987).
6. A. A. Gavrilov, Exhalational-Sedimentary Manganese Ore Accumulation [in Russian], Nedra, Moscow (1972).

7. E. G. Gurvich, Yu. A. Bogdanov, and A. P. Lisitsyn, "Types of hydrothermal formations on the ocean floor and their evolution," in: *The Evolution of Sedimentary Ore Formation during the Earth's History* [in Russian], Nauka, Moscow (1984), pp. 41-65.
8. K. K. Zelenov, "Iron and manganese in the exhalations of the submarine volcano Banu-Wuhu (Indonesia)," *Dokl. Akad. Nauk SSSR*, 155, No. 6 (1964).
9. B. P. Zolotarve, "The petrology of the basalts of the modern ocean in connection with their tectonic position," *Geotektonika*, No. 1, 22-35 (1979).
10. D. Kronen, *Submarine Mineral Deposits* [Russian translation], Mir, Moscow (1982).
11. N. A. Lisitsyn, *The Removal of Chemical Elements during the Weathering of Basic Rocks* [in Russian], Nauka, Moscow (1973).
12. A. P. Lisitsyn, *Processes of Oceanic Sedimentation* [in Russian], Nauka, Moscow (1978).
13. A. P. Lisitsyn, Yu. A. Bodanov, I. O. Murdmaa, et al., "Metalliferous sediments and their genesis," in: *Geological-Geophysical Investigations of the Southeastern Portion of the Pacific Ocean* [in Russian], Nauka, Moscow (1976).
14. A. P. Lisitsyn and Yu. A. Bogdanov (eds.), *Metalliferous Sediments of the Red Sea* [in Russian], Nauka, Moscow (1986).
15. A. P. Lisitsyn and E. G. Gurvin (eds.), *Metalliferous Sediments of the Indian Ocean* [in Russian], Nauka, Moscow (1987).
16. A. A. Migdisov, V. A. Bogdanov, A. P. Lisitsyn, et al., "The geochemistry of metalliferous sediments," *Metalliferous Sediments of the Southeastern Portion of the Pacific Ocean* [in Russian], Nauka, Moscow (1979), pp. 122-200.
17. T. V. Rozanova and G. N. Baturin, "Ore hydrothermal shows on the floor of the Indian Ocean," *Okeanologiya*, 11, No. 6, 874-879 (1971).
18. P. Rona, *Hydrothermal Mineralization in Spreading Regions of the Ocean* [Russian translation], Mir, Moscow (1986).
19. A. B. Ronov and A. A. Yaroshevskii, "The chemical make-up of the Earth's crust," *Geokhimiya*, No. 11, 1652-1678 (1967).
20. N. M. Strakhov, *Basis for the Theory of Lithogenesis* [in Russian], Izd. Akad. Nauk SSSR (1962).
21. N. M. Strakhov, "Mid-oceanic risulfides as sources of ore components," *Litol. Polezn. Iskop.*, No. 3, 20-37 (1974).
22. N. M. Strakhov, *Problems in the Geochemistry of Modern Oceanic Lithogenesis* [in Russian], Nauka, Moscow (1976).
23. L. E. Shterenberg, "Modern ore formation on the floors of the oceans," *Geologiya Rud. Mest.*, No. 4, 76-91 (1984).
24. G. C. Alt, P. Lonsdale, R. Haymor, and K. Muchlenbachs, "Hala sulfide and oxide deposits on seamounts near 21°N East Pacific Rise," *Bull. Geol. Soc. Am.*, 98, 2, 157-68 (1987).
25. E. Bonatti, "The origins of metal deposits in the oceanic lithosphere," *Sci. Am.*, 238, 54-68 (1978).
26. E. Bonatti, "Hydrothermal metal deposits from the oceanic rifts: a classification," *Hydrothermal Process. Seafloor Spread. Cent. Proc.*, NATO ADV. Res. Inst., Cambridge, April 5-8 (1982), N.Y.-London (1983), pp. 491-502.
27. E. Bonatti, G. Honnorez, and M. Honnorez-Guerstein, "Copper-iron sulfide mineralization in the equatorial Mid-Atlantic," *Econ. Geol.*, 71, 8, 1515-1525 (1976).
28. K. Bostrom and N. M. Peterson, "The origin of aluminum-poor ferromanganoan sediments in area of heat flow on the East Pacific Rise," *Marine Geol.*, 7, 5, 427-448 (1969).
29. G. R. Cann, S. C. Winter, and R. G. Pritchard, "A hydrothermal deposit from the floor of the Gulf of Aden," *Mining Mag.*, 41, 193-199 (1977).
30. D. S. Cronan, "Metalliferous at ocean spreading centers — a conference report," *J. Geol. Soc. London*, 136, 621-626 (1976).
31. G. Dymond, G. B. Corliss, G. R. Heath, et al., "Origin of metalliferous sediments from the Pacific Ocean," *Bull. Geol. Soc. Am.*, 84, 3355-3372 (1973).
32. G. M. Edmond, C. Measures, B. Mangum, et al., "On the formation of metal-rich deposits at ridge crests," *Earth Planet Sci. Lett.*, 46, 19-30 (1979).
33. R. M. Haymon, "Hydrothermal deposition on the East Pacific Rise at 21°N," *J. Geochem. Explor.*, 19, 1-3, 493-395 (1983).
34. R. M. Haymon and M. Kastner, "Hot spring deposits on the East Pacific Rise at 21°N. Preliminary description of mineralogy and genesis," *Eath Planet Sci. Lett.*, 53, 3, 363-381 (1981).
35. R. Hekinian, M. Fevrier, J. Bischoff, and W. Shanks, "Sulfide deposits from the East Pacific Rise near 21°N," *Science*, No. 207, 1433-1444 (1980).

36. R. Hekinian and G. Fouquet, "Volcanism and metallogenesis of axial and off-axial structures on the East Pacific Rise near 13°N," *Econ. Geol.*, 80, 1, 221-255 (1985).
37. M. Hoffert, A. Perseil, R. Hekinian, et al., "Hydrothermal deposits sampled by diving saucer in transform fault "A" near 37°N on the Mid-Atlantic Ridge, FAMOUS area," *Oceanol. Acta*, 1, 73-86 (1978).
38. M. Hoffert, A. Person, C. Courtois, et al., "Sedimentology, mineralogy, and geochemistry of hydrothermal deposits from holes 424, 424A, 424B, and 424C (Galapagos spreading center)," *Initial Reports of DSDP, Wash.*, 54, 339-376 (1980).
39. P. F. Lonsdale, G. H. Bischoff, V. M. Burns, et al., "High-temperature hydrothermal deposits in seabed at the Gulf of California spreading center," *Earth Planet Sci. Lett.*, 49, 8-20 (1980).
40. A. Malahoff, "A comparison of the massive submarine polymetallic sulfides of the Galapagos rift with some continental deposits," *Marine Technol. Soc. J.*, 16, 3, 39-45 (1982).
41. M. A. McKibben and W. A. Elders, "Fe-Zn-Cu-Pb mineralization in the Salton Sea geothermal system, Imperial Valley, California," *Econ. Geol.*, 80, No. 3, 539-559 (1985).
42. E. Oudin, "Hydrothermal sulfide deposits of the East Pacific Rise (21°N), Pt. 1," *Marine Mining*, 4, No. 1, 39-72 (1983).
43. P. A. Rona, "Criteria for recognition of hydrothermal mineral deposits in oceanic crust," *Econ. Geol.*, 73, 2, 135-161 (1978).
44. M. R. Scott, R. B. Scott, P. A. Rona, et al., "Rapidly accumulating manganese deposits from the median valley of the Mid-Atlantic Ridge," *Geophys. Res. Lett.*, 1, 335-358 (1974).
45. S. Shearme, D. S. Cronan, and P. A. Rona, "Geochemistry of sediments from the TAG hydrothermal field, M.A.R. at latitude 26°N," *Marine Geol.*, 51, Nos. 3-4, 269-293 (1983).
46. G. R. Toth, "Deposition of submarine crusts rich in manganese and iron," *Bull. Geol. Soc. Am.*, 91, 44-54 (1980).
47. K. Turekian and K. H. Wedephol, "Distribution of the trace elements in some major units of the Earth's crust," *Bull. Geol. Soc. Am.*, 72, No. 2 (1961).
48. I. M. Varentsov, B. A. Sakharov, V. R. Drits, et al., "Hydrothermal deposits of the Galapagos rift zone, Leg 70: mineralogy and geochemistry of major components," *Initial Reports of the DSDP, Wash.*, 70, 235-268 (1983).

FERROMANGANESE CRUSTS IN THE ATLANTIC:
GEOCHEMISTRY OF RARE-EARTH ELEMENTS AND ASPECTS
OF ORIGIN, SUBMARINE SEAMOUNT KRYLOV

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Data on the distribution of REE in ferromanganese oxide-hydroxide crusts on the submarine seamount Krylov (Cape Verde abyssal platform, eastern Atlantic) are presented. The behavior of Σ REE, heavy and light REE, Ce, and Eu suggests that hydrothermal components were significant in the early stages of formation of oxide-hydroxide crusts, and that later, during subsidence and eastward movement of the platform, hydrogenic factors, developed by means of a sorption mechanism, became dominant.

Data from observations in the field and from experimental studies have shown that the formation of ferromanganese crusts, unlike nodules, is associated with the growth of hydroxide phases during unidimensional entry of accumulating components at the solution-substrate interface [1, 4]. The geochemical environment of mineral formation is accurately recorded in the composition of crustal buildups. Ferromanganese crusts differ between hydrothermal and hydrogenic deposits. These differences appear quite clearly, along with other characteristics, in the distribution of rare earth elements (REE). The accumulation of hydrothermal components on the sea floor does not take place in isolation from seawater, however. This is shown by direct observation and by data from underwater sampling: significant dilution with seawater takes place directly adjacent to hydrothermal discharges [21, 32]. Moreover, sediments formed under such circumstances may interact with seawater, which commonly leads to an appreciable change in their original composition.

We have previously presented data from a detailed study of the mineralogy and crystallochemical characteristics of ferromanganese oxide-hydroxide phases in crusts and substrate and the geochemistry of the principal components and heavy metals [2]. The problem in that investigation was to determine, on the basis of a study of the REE geochemistry, the genetic features of ferromanganese oxide-hydroxide crusts in the area of the submarine seamount Krylov, particularly the relationship between hydrothermal and hydrogenic components during their geochemical evolution and deposition.

The basis of the work was a study of samples of ferromanganese oxide-hydroxide crusts and the volcanic substrate, obtained on the 1st voyage of the research vessel "Academician Nikolai Strakhov," which was engaged in a series of geologic and geographic studies in the area of the Krylov seamount (Cape Verde, eastern Atlantic). A preliminary description of the material dredged up was made on board by B. P. Zolotarev, V. A. Evroshchev-Shak, and I. I. Bebeshev.

A detailed description of the sampling methods and the methods of study of the mineral composition, content of principal components, and heavy metals has been presented in [2, 31]. It must be noted that identification of the finely dispersed ferromanganese oxide-hydroxide phases was accomplished using precision x-ray diffractometry and electron microdiffraction, accompanied by microprobe determination of the chemical composition of individual micro-particles, by means of the x-ray energy dispersive apparatus "Kevex" and a transmitting scanning electron microscope.

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Geological Data and Mineralogy. The submarine seamount Krylov is a volcanic structure of the central type, located in the Cape Verde abyssal plain. Its formation is related to the extrusion of subalkaline to alkaline, slightly differentiated basalt along an east-west fault [3].

The geochemistry of the REE was studied in samples of ferromanganese oxide-hydroxide crusts and substrate (1-18-D-1-54; 1-18-D-54-2), brought up from station 18 on the middle reaches of the slopes of the mountain (17°32.8' N, 29°59.71'W, depth 2737 m).

The general characteristics and mineral content of components and layers forming the ferromanganese crusts are briefly described in the note at the bottom of Table 1. It is important to note that if the main components in the upper 5-7 millimeters of layer 1 (see Table 1) are Fe-vernadite and Mn-feroxyhyte ($\text{Mn-}\delta'\text{-FeOOH}$) with lesser amounts of birnessite and minor asbolane-buserite and goethite, then the Mn-feroxyhyte content will diminish markedly toward the base of the crust. In the lower layers 6 and 7 (see Table 1), Mn-feroxyhyte is absent and the main Fe oxide-hydroxide is goethite. The widespread development of goethite at the base of the crust is caused by alteration of Mn-feroxyhyte under oxidizing conditions, where there are small changes in the Mn-Fe ratio (Mn/Fe 0.64-0.89, rarely to 1.13) through the crust, and by the significant effect of hydrothermal components.

Crustal hydroxide growths that form on the substrate represent an almost complete hydrothermal alteration of hyaloclastites in alkaline basalt to Fe-palygorskite.

Distribution of REE and Origin of Ferromanganese Crusts. Information on REE in ferromanganese oxide-hydroxide crusts and rocks of the substrate is presented in Table 1 and is graphically illustrated in Figs. 1 and 2. A number of essential points can be seen from an analysis of these data.

REE Distribution. REE in ferromanganese crusts, compared to the standard for shale (see Fig. 1), are characterized by a distribution that is about average, close to that of relatively shallow-water (2150 m) varieties, or to the average composition of oxide-hydroxide deposits on seamounts in the Line Islands archipelago of the central Pacific [5], and a clearly expressed positive cerium anomaly (Ce/Ce^* 2.3-4.1; see Table 1). It must be emphasized, however, that crusts developed at great depths (>2150 m and especially at 4700 m) in the Line Islands region are characterized by depletion in total REE (about 2-3 times, compared to shallow-water varieties) and a negative value for the cerium anomaly (Ce/Ce^* 0.96-0.51). These data may indicate a significant role of seawater at different levels as solvents controlling the composition and concentration of REE in submarine crusts. This conclusion is supported by the REE distribution in the water column of both the Atlantic and Pacific Oceans [10-12]: the Ce concentration is sharply decreased with depth.

A decrease in ΣREE ($\mu\cdot 10^{-4}\%$) from surface to basal layer was observed in the crusts studied: 1913.1 $\rightarrow$ 2093.3 $\rightarrow$ 1453.4 (see Table 1). A direct relationship is clearly shown between the lanthanide content and (Mn + Fe) (Fig. 3A, $r=0.74$). This reflects the dominance of the sorption mechanism for accumulation of REE by oxide-hydroxides of Mn and Fe [5, 9, 27]. Table 1 shows a tendency for an increase in the value $\Sigma(\text{La-Lu})/(\text{Mn} + \text{Fe})$ and especially $\Sigma(\text{La-Lu})/\text{Fe}$ from the basal layer to the surface of the crustal deposits. This distribution of these ratios, considering the known persistence of Mn/Fe in the crustal section, can be interpreted as an indicator of the importance of hydrogenic processes on the accumulation of the later crustal layers.

The distribution of light and heavy REE in the section of the ferromanganese crust is shown by a number of ratios (see Table 1). If the behavior of the ratios $\Sigma(\text{La-Sm})/\Sigma(\text{Eu-Lu})$ and Yb/Sm shows a clear trend toward a decrease in the light REE fraction (with a corresponding increase in the heavy lanthanides) from the basal layer to the surface, then a less clear picture is seen in the values of La/Yb and La/Sm (see Fig. 3A).

It is very important, in the interpretation of these data, that a negative relationship is seen between the values of Fe and La/Sm ($r = -0.45$; see Fig. 3B). A close correlation is also established between Mn/Fe-La/Sm : for sample 1-18-D-54-1, $r = 0.7$; for sample 1-18-D-54-2, $r = 1.0$; for the crust as a whole at this station, $r = 0.8$ (see Fig. 3C). In other words, these data indicate that oxide-hydroxides of manganese, mainly in Fe-vernadite, are the chief carriers of light rare earths (LREE).

Correspondingly, the heavy fractions (HREE) are associated with oxide-hydroxides of iron (see Fig. 3D): there is a rather high correlation ($r = 0.89$) between the magnitudes of Fe

TABLE 1. REE Content and Components of Ferromanganese Crusts (submarine seamount Krylov, Eastern Atlantic), $\cdot 10^{-4} \%$

Component	Sample number							Sample number								
	1-18-D-54-1							1-18-D-54-2				z-55	z-56	z-57	z-92	average
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
La	291	297	269	180	150	148	75	285	83	54	286	10	11	19	21	15,2
Ce	1275	1440	1390	1210	1336	1133	93,0	1420	440	193	1600	20	20	29	27	24,0
Nd	258	255	278	188	176	126	113	289	72	57	267	-	-	-	-	-
Sm	45,7	47,8	46,0	27,6	27,2	24,3	62,9	49,1	15,4	10,8	46,3	4,6	4,4	6,0	6,0	5,2
Eu	11,9	13,4	13,4	7,9	8,8	7,2	4,6	14,1	4,1	2,6	12,1	1,6	1,4	1,9	1,8	1,7
Tb	9,5	11,6	9,5	5,5	5,9	4,9	3,3	10,1	3,3	1,9	9,4	0,83	0,80	1,1	0,93	0,91
Yb	19,1	25,1	20,6	11,0	13,2	8,2	7,5	20,8	8,9	5,2	16,8	2,5	2,4	3,2	2,7	2,7
Lu	2,9	3,4	3,1	2,1	2,4	1,8	0,9	3,3	1,1	0,69	2,81	0,38	0,41	0,52	0,41	0,43
Sc	9,46	14,0	14,1	16,6	12,9	14,9	26,6	10,2	14,8	19,8	7,42	56	53	56	39	51,0
$\Sigma \text{ REE (La-Lu)}$	1913,1	2093,3	2029,6	1632,1	1719,5	1453,4	360,2	2091,4	627,8	325,19	2240,41	39,91	40,41	60,72	59,84	50,14
Ce/Ce**	2,33	2,61	2,54	3,29	4,08	4,15	0,49	2,47	2,84	1,74	2,89	-	-	-	-	-
$\Sigma (\text{La-Sm})/\Sigma (\text{Eu-Lu})$	43,08	38,13	42,55	60,59	55,75	64,76	21,10	42,30	35,08	30,30	53,50	6,51	7,07	8,04	9,25	7,73
La/Yb	15,24	11,83	13,06	16,36	11,36	18,05	10,00	13,70	9,33	10,38	17,02	4,00	4,58	5,94	7,78	5,63
La_N/Yb_N	1,33	1,03	1,14	1,43	0,99	1,57	0,87	1,19	0,81	0,90	1,48	0,36	0,39	0,60	0,68	0,49
La/Sm	6,37	6,21	5,85	6,52	5,51	6,09	1,19	5,80	5,39	5,00	6,18	2,17	2,5	3,17	3,5	2,92
Ce/La	4,38	4,85	5,17	6,72	8,91	7,65	1,24	4,98	5,30	3,57	5,59	2,00	1,82	1,53	1,29	1,58
Ce/Sm	27,90	30,12	30,22	43,84	49,12	46,62	1,48	28,92	28,57	17,87	34,56	4,35	4,54	4,83	4,50	4,62
Eu/Sm	0,26	0,28	0,29	0,29	0,32	0,30	0,07	0,29	0,27	0,24	0,26	0,35	0,32	0,30	0,33	0,25
Nd/La	0,89	0,86	1,03	1,04	1,17	0,85	1,51	1,01	0,87	1,06	0,93	-	-	-	-	-
Ce/Yb	66,75	57,37	67,48	110,00	101,21	138,17	12,40	68,27	49,44	37,11	95,24	8,00	8,33	9,06	10,00	8,89
Yb/Sm	0,42	0,53	0,45	0,40	0,49	0,34	0,12	0,42	0,58	0,48	0,36	0,54	0,55	0,53	0,45	0,52
Fe _{tot}	19,371	21,156	21,480	19,189	21,630	19,481	4,021	17,765	10,365	4,385	13,918	-	-	-	-	-
Mn _{tot}	16,703	13,537	16,347	17,175	15,248	15,378	1,293	15,534	4,644	0,89	16,324	-	-	-	-	-
Mn-Fe	0,862	0,640	0,761	0,895	0,705	0,789	0,322	0,874	0,448	0,203	1,156	-	-	-	-	-

Note: Air-dried weight, neutron activation analysis. Components of crustlike layers of ferromanganese oxide-hydroxides and hydrothermally altered substrate: palygorskitized hyaloclastite of alkali basalt (sample 1-18-D-54-1): 1) upper layer of black ferromanganese oxide-hydroxides; mainly Fe-vernadite and Mn-feroxyhyte ($\text{Mn-}\delta^{\text{I}}\text{-FeOOH}$), with lesser amounts of birnessite, and minor asbolane-buserite and goethite (see Fig. 2); 2) layer of black ferromanganese oxide-hydroxide (25-30 mm). Main components: Fe-vernadite, Mn-feroxyhyte, and goethite, with lesser amounts of birnessite, and minor asbolane-buserite, quartz, zeolites (lomonite, mordenite, heulandite), and a mixed-layer mica phase: smectite, smectite and palygorskite; 3) layer of black, mainly ferromanganese oxide-hydroxides (3-7 mm); mostly Fe-vernadite, Mn-feroxyhyte, and goethite, with traces of smectite; 4) layer of black, mainly ferromanganese oxide-hydroxides (2-7 mm); 5) brownish-black layer, highly impregnated with ferromanganese oxide-hydroxides (up to 35 mm); mostly goethite with lesser amounts of Fe-vernadite and apatite; 6) basal layer of black, mainly ferromanganese oxide-hydroxides (2-10 mm). Main components: goethite and Fe-vernadite, with lesser amounts of asbolane-buserite and minor todorokite; 7) substrate, hyaloclastics of alkaline basalt, highly altered to Fe-palygorskite. Composition of Mn-Fe oxide-hydroxide layers in crusts and hydrothermally palygorskitized substrate, hyaloclastite in alkaline basalt (sample 1-18-D-54-2): 8) uppermost crust of Mn-Fe oxide-hydroxides (7-10 mm); mostly Fe-vernadite and x-ray amorphous Mn-Fe oxide-hydroxides (possibly Mn-feroxyhyte?); 9) Mn-Fe oxide-hydroxides in an intensely mottled zone developed in highly reworked hyaloclastites (10-15 mm); material is x-ray amorphous Mn-Fe oxide-hydroxides (possibly Fe-vernadite and Mn-feroxyhyte), palygorskite with a minor amount of mica, smectite, hydrogoethite, and quartz; 10) cream-colored, hydrothermally reworked hyaloclastitic substrate; palygorskite with minor mica (?), smectite, and hydrogoethite; 11) sooty black Mn-Fe oxide-hydroxides forming lateral crusts; material is Fe-vernadite and x-ray amorphous Mn-Fe oxide-hydroxides (Mn-feroxyhyte?), with trace amounts of apatite. Alkaline basalts from submarine seamount Krylov (from data of B. P. Zolotarev): 12-15) olivine basalt partly, poorly, and slightly altered hydrothermally, respectively; 16) hyalobasalt, very poorly altered hydrothermally; average value (samples 12-15).

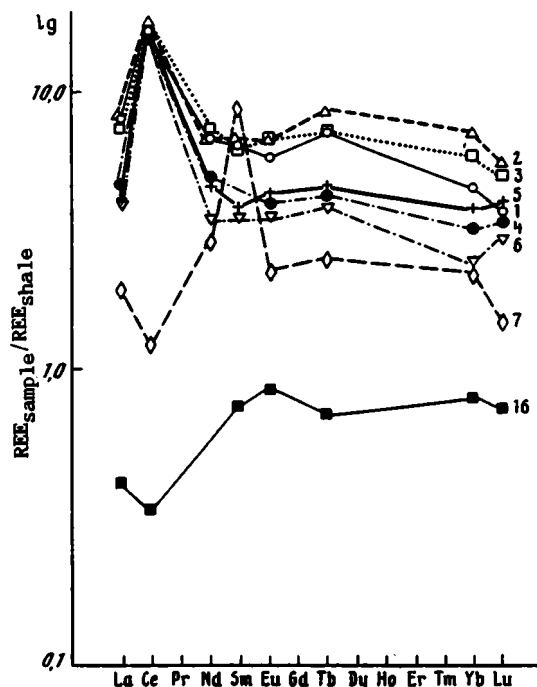


Fig. 1. REE distribution compared to standard for shale [16] in ferromanganese oxide-hydroxide deposits and in substate, hydrothermally altered (palygorskitized) hyaloclastite in alkaline basalt (submarine seamount Krylov, Cape Verde Basin, eastern Atlantic. Component (1-7, 16) listed in footnote to Table 1.

and Yb/Sm, which is in accord with the negative relationship between Mn/Fe and Yb/Sm ($r = -0.73$).

The results presented may be interpreted in light of the sorption mechanism of REE, taking into account the various physicochemical characteristics of the surface of Mn and Fe oxide-hydroxides, the form of RE found in seawater, and possible changes in the mixture of seawater and hydrothermal solutions at various stages in the accumulation of the layers making up the crusts.

In the case of vernadite (or, as referred to in the foreign literature, δMnO_2 or $2.4 \text{ \AA MnO}_2$) and birnessite it has been established that the pH neutral charge point [pH(NCP)] for these compounds as a rule is less than 3 (from 1.4 to 2.25). The charge on the surface, determined by titration, is considerably higher than for other oxide-hydroxides: δMnO_2 , $27.2 \pm 2.2 \text{ mol/kg}$; αFeOOH , 1.3 mol/kg ; $\text{Fe}(\text{OH})_3$, 10.9 mol/kg , while the specific surface can vary from 74 to $350 \text{ m}^2/\text{g}$, depending on the environment of synthesis of these phases and conditions during their aging [6, 7]. The low pH(NCP) value for these modifications of MnO_2 can be used to explain why a number of anions (Cl^- , SO_4^{2-} , etc.) are not absorbed by these phases in the pH interval from 2 to 9 [6]. In contrast to MnO_2 , such marine modifications of Fe oxide-hydroxides as $\alpha\text{-FeOOH}$ (goethite) are characterized by a pH(NCP) of 7.5 [6, 7]. These investigators have shown that the sorption capacity of these varieties of MnO_2 for Na, Mg, Ca, and K in seawater is almost an order of magnitude greater than in goethite.

In other words, in interpreting these data on the behavior of REE it is necessary to keep in mind that such significant differences in surface properties mean that ions with the properties of cations are absorbed mainly by oxide-hydroxides of Mn, while anions or complex ions with a relatively low charge density are absorbed by oxide-hydroxides of Fe.

Results of a study of the form of occurrence of REE in seawater [29] and data on natural deposits [11, 12] suggest that light REE are present chiefly in ion form, while heavy REE occur in complex compounds. For example, the proportion (%) of occurrence of La^{3+} in the form of free ions is 2 times greater than for Nd^{3+} , and the relative amount of the ion form decreases as the REE weight increases. Thus, differences in the physico-chemical properties of the surface of oxide-hydroxides of Mn and Fe and the form of light and heavy REE show up,

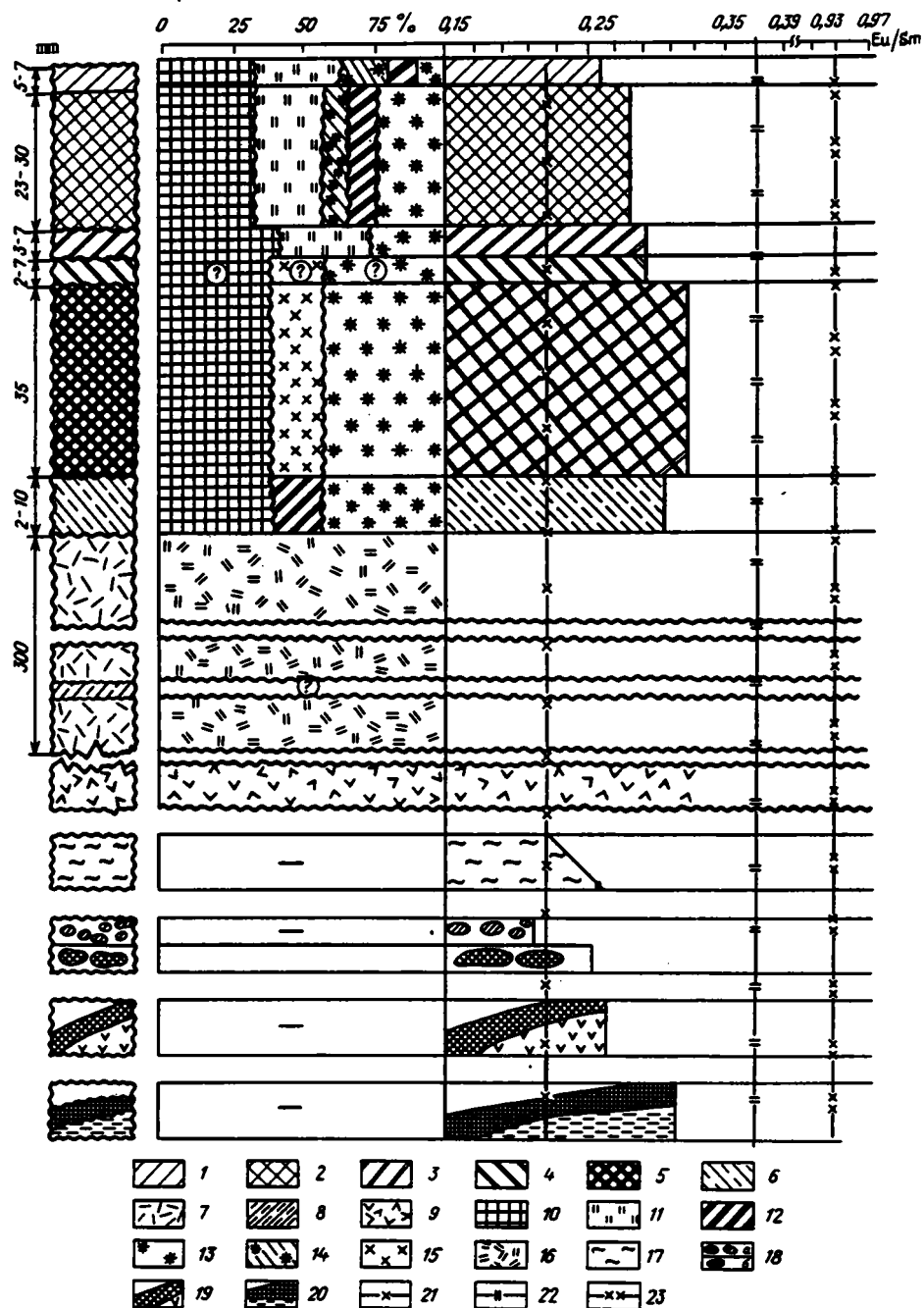


Fig. 2. Composition, mineralogy, and distribution of the value Eu/Sm through the section of ferromanganese oxide-hydroxide crust deposits and substrate, hydrothermally altered (palygorskitized) hyaloclastite in alkaline basalt (sample 1-18-D-54-1). 1-7) Components (see Table 1); 8) veins and nodules of Mn-Fe hydroxides in substrate; 10) Fe-vernadite; 11) Mn-feroxyhyte; 12) asbolane-buserite; 13) goethite; 14) birnessite; 15) apatite; 16) Fe-palygorskite; 17) seawater (average) [11]; 18) shallow-water ferromanganese concretions; 19) vernaditic ferromanganese crusts (average at depths of 1 and 4.7 km), submarine seamounts in the Line Islands archipelago (NW Pacific [5]); 20) Mn crusts in hydrothermal rocks, composed of todorokite (Galapagos rift zone [22]); 21-23) Eu/Sm values (21, average for shale [16]; 22, tholeiitic basalt on the axis of the East Pacific Rise [21]; 23, sample of solution from near a submarine hydrothermal source, composed of a mixture of 5% hydrothermal fluids and 95% seawater, East Pacific Rise [21]).

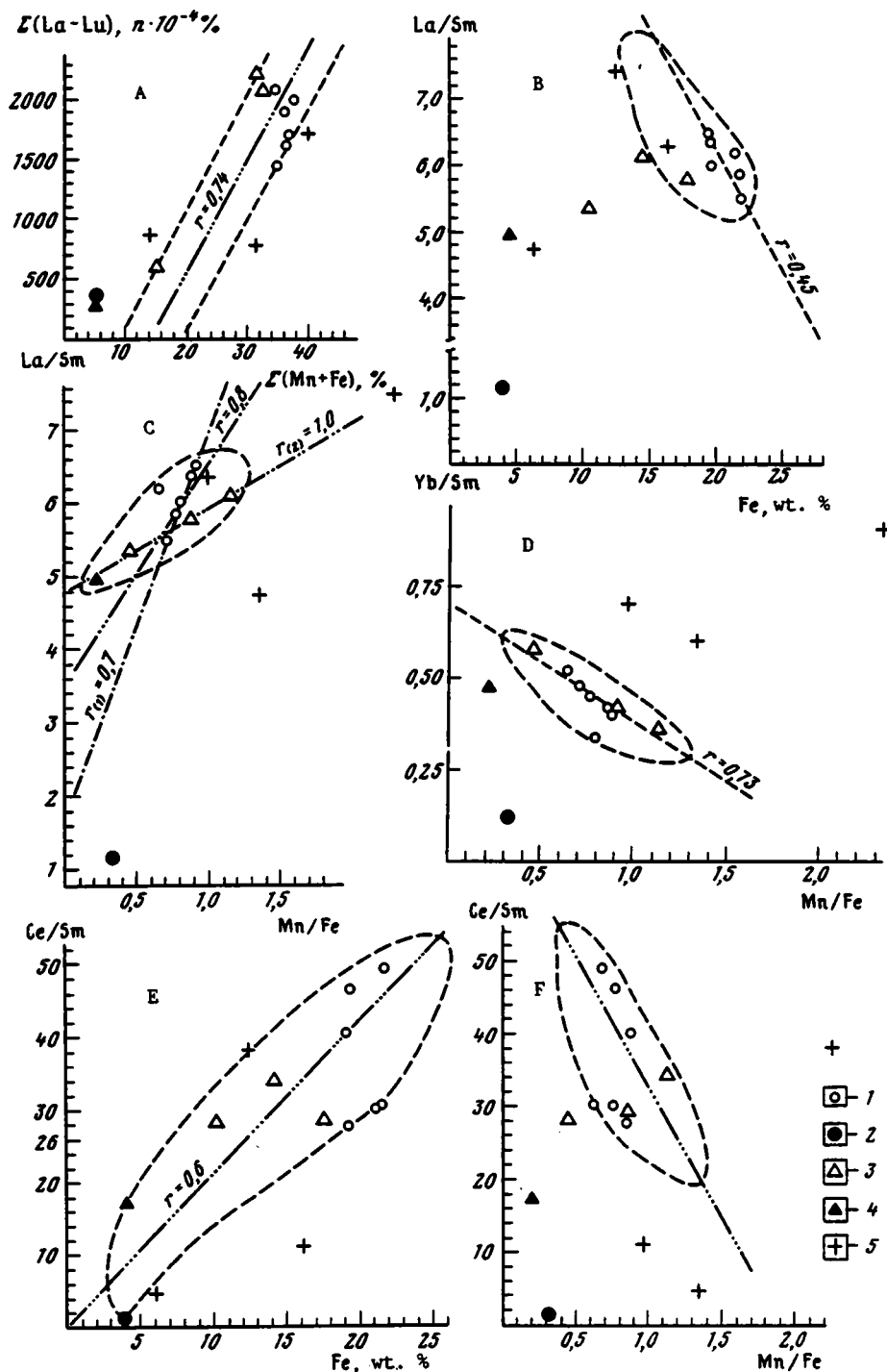


Fig. 3. Correlation diagram for REE and the main components of ferromanganese oxide-hydroxide deposits and substrate-hydrothermally altered (palygorskitized) alkaline basalt hyaloclastites (Cape Verde basin). A) $\Sigma(\text{Mn} + \text{Fe}) - \Sigma(\text{La-Lu})$; B) $\text{Fe} - \text{La/Sm}$; C) $\text{Mn/Fe} - \text{La/Sm}$; D) $\text{Mn/Fe} - \text{Yb/Sm}$; E) $\text{Fe} - \text{Ce/Sm}$; F) $\text{Mn/Fe} - \text{Ce/Sm}$; 1-2) sample 1-18-D-54-1 (1, layer of ferromanganese deposits; 2, substrate); 3-4) sample 1-18-D-54-2 (3, layer of ferromanganese deposits; 4, substrate); 5) Mn-Fe crust of vernaditic composition, seamount in the Line Islands archipelago region (NW Pacific [5]).

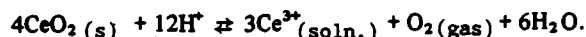
as shown above (see Fig. 3A-E), in the preferential accumulation of LREE on oxide-hydroxides of Mn and of HREE on oxide-hydroxides of Fe.

The observed decrease in the value of $\Sigma(\text{La-Sm})/\Sigma(\text{E-Lu})$ from the base of the crusts to the surface, at the same time that the value of Mn/Fe persists through the section (see Table 1), may indicate a relative decrease in the proportion of hydrothermal components in the seawater-hydrothermal solution mix. It has been established in [21], which was devoted to the study of REE in hydrothermal sources on the East Pacific rise, that such solutions are considerably enriched in LREE compared to seawater.

Features of the behavior of Ce during accumulation in oxide-hydroxide crusts may be illustrated by the example of the distribution of the ratios Ce/Ce*, Ce/La, Ce/Yb, and Ce/Sm (see Table 1), which systematically decrease from the basal layers to the top. Calculation of the Ce anomaly was accomplished according to the expression $\text{Ce}/\text{Ce}^* = .2(\text{Ce}/\text{Ce}_{\text{sh}})/(\text{La}/\text{La}_{\text{sh}} + \text{Nd}/\text{Nd}_{\text{sh}})$, where Ce, La, and Nd are the content in the sample and Ce_{sh}, La_{sh}, and Nd_{sh} are the concentration in shale [16]. On the correlation diagram (see Fig. 3E), Fe-Ce/Sm clearly shows a direct relationship between these values ($r = 0.6$), which puts Ce in a group with heavy REE. However, the relationship between Ce/Sm and Mn/Fe is less clearly a negative one (see Fig. 3F), which is consistent with the behavior of Ce noted above.

On the whole, the high positive value of the cerium anomaly (Ce/Ce*) and the magnitudes of Ce/La, Ce/Yb, and Ce/Sm for ferromanganese oxide-hydroxides in marine deposits can be seen as indicators of hydrogenic effects [14, 15, 23-26]. The truth of this view is particularly well shown by the example of hydrothermal ferromanganese oxide-hydroxide deposits, using the difference in transport distance from their place of precipitation from solution, the axial zone of the East Pacific Rise [27]. If Ce/Sm = 0.33 within 9 km of the axis of the rise, then during transportation over 1085-1139 km the magnitude of this ratio increases correspondingly to 1.83-1.84. The observed increase $\Sigma\text{REE} (\cdot 10^{-4})$ over the same distance is 5-7 times (from 103.96 to 787.43).

Along with this, an analysis of data on REE in oxide-hydroxide crusts on seamounts in the Line Islands region, central Pacific [5], suggests that variations in the relationships under discussion are controlled to a considerable degree by ocean depth. Comparison of values for three representative depth intervals, sample B (1150-1325 m), sample I (2150-2650 m), and sample S (4700 m), shows a definite tendency toward a decrease with depth: Ce/Ce* (B), 3.01; (I), 0.96; (S), 0.51; Ce/La (B), 5.13; (I), 1.73; (S), 0.96; Ce/Yb (B), 42.33; (I), 15.68; (S), 7.58; Ce/Sm (B), 38.36; (I), 11.04; (S) 4.58. These results may be interpreted in light of data on the distribution of REE, and especially Ce, in the water column of the Pacific and Atlantic Oceans. Studies by DeBaar and others [10, 11] have shown that the Ce concentration in the Atlantic is about 2 times higher than in the Pacific, and with an increase in depth in both oceans there is typically a depletion in the element. In the Atlantic, the maximum value of Ce, particularly apparent in the distribution of Ce/Ce*, is recorded in the top layer (100-250 m) and diminishes gradually with depth inversely with the content of dissolved O₂. It is widely recognized that the behavior of Ce in the oceanic water column is controlled primarily by oxidation-reduction reactions of the type [11]



This reaction is effected by means of sorption interaction between phases (in this case with Fe oxide-hydroxides) with a strong part played by cyclical biochemical processes and circulation of the water masses. In this aspect, the geochemical features of the behavior of Ce in the oceanic water column are close to those of transition 3d-metals: Mn, Fe, Co, and others [1, 9, 19, 30, 33].

Thus, upon an examination of the clear-cut decrease in the values of Ce/Ce*, Ce/La, Ce/Yb, and Ce/Sm (see Table 1) from the basal layers of hydroxide crusts to their surface in the context of the overall behavioral patterns of REE discussed above, as well as the heavy and light REE fractions, it can be concluded that hydrogenic factors are dominant. During the early stages of formation of crustal deposits on the Krylov seamount, the volcanic structure was located considerably closer to the axial zone of the Mid-Atlantic ridge, where hydrothermal components obviously were more important and the depth below sea level was relatively small. This is also indicated by the presence of hyaloclastites, which form the substrate, and lava flow breccia near the top. In just such shallow, well aerated conditions, in which seawater is distinguished by its highest Ce concentration, were the basal ferromanganese layers of the crust formed, with a high value of the Ce anomaly (Ce/Ce* = 4.15; see Table 1). During subsequent translation of the Cape Verde plain to the east and

its subsidence to abyssal depths, the Ce/Ce* value for the upper layers decreased to 2.33. However, like a mantle diapir, during its growth the Krylov seamount may also have undergone a certain amount of bulging upward from the abyssal floor of the plate [3].

Specific behavior of Eu is graphically described in the distribution of Eu/Sm values (see Fig. 2 and Table 1): the magnitude of this ratio decreases systematically from the lower, basal part of the ferromanganese oxide-hydroxide crust (0.30-0.32) to the surface (0.26), while remaining at remarkably high level, characteristic for shale (0.21) and accordingly continental run-off, as well as for seawater (0.26).

Such variations may indicate an appreciable role for hydrothermal processes in the early stage of formation of crustal deposits. For example, in hydrothermal Mn oxide-hydroxide (todorokite) deposits, crusts of the Galapagos rift zone [21, 31] have Eu/Sm = 0.31, which is close to magnitudes noted above for the basal layers of crusts on the seamount Krylov.

According to data presented in [21], samples of hydrothermal solutions taken by means of the submersible "Syana" in the region of 13° North on the East Pacific rise have an REE concentration approximately 1000 times higher than the seawater, with an Eu/Sm value of about 2.6 (in seawater of the open ocean it is 0.26). Contamination of these samples with seawater ranged from 5 to 95%. In order to estimate the amount of seawater, the authors assumed that the Mg content in the end member of the series hydrothermal solution - seawater was zero. If the magnitude of Eu/Sm in relatively pure hydrothermal solution (5-50% seawater) ranged from 2.19 to 3.18, then with considerable dilution by seawater (95%) this ratio falls to 0.95. It is important to note that in [21] it was indicated that there was not a strong proportionality between the REE content and contamination of hydrothermal solutions by seawater, which is possibly due to insufficient care in taking the samples or to mistakes made during the storage and analysis of the solutions. Nevertheless, the persistence of this trend is quite definite. There is reason to believe that the magnitude of Eu/Sm in ferromanganese oxide-hydroxide nodules, deposits, and associated sediments reflects the value of this ratio in the solutions. For example, in shallow-water ferromanganese nodules the magnitude of Eu/Sm is 0.21, which corresponds to the value of the ratio in river water and in seawater in shallow regions of the ocean, and is close to the value for shale [16]. This conclusion is also correct for ferromanganese oxide-hydroxide nodules in various zones of the ocean: Eu/Sm = 0.20-0.25 [25]. Thus, despite individual rare exceptions, it is fair to consider that this trend reflects the original properties of the depositional environment of ferromanganese oxide-hydroxide sediments and suspensions. This view can be illustrated by the example of the Eu/Sm distribution in hydrothermal metalliferous deposits [27], which have been deposited both in the immediate axial zone of the East Pacific Rise (9 km from the axis, Eu/Sm is 0.26) and upon transport for considerable distances (1085 and 1139 km from the axis, Eu/Sm is 0.25 and 0.27, respectively). It is significant that along the extent of this traverse the value of Eu/Sm is characterized by persistence. Extrapolating the relationship cited above to the relatively ancient beds of the ferromanganese crusts on seamount Krylov (Eu/Sm = 0.30-0.32) suggests that the ore-forming solution was primarily seawater in which hydrothermal components formed no more than 1-2% in the early stages of deposition (beds 3-6). Higher parts of the crust (Eu/Sm = 0.26-0.28; beds 1, 2; see Fig. 2 and Table 1) were formed during considerable dilution of the hydrothermal source with seawater, which left a hydrogenic imprint on the ferromanganese deposits. However, if a 1-2% content of hydrothermal components mixed with seawater has a relatively mild effect on the REE balance, especially Eu and Sm, then such a mixture may have a considerable effect on the distribution of Mn and Fe, for which the content of dissolved and suspended forms of these metals is several orders of magnitude greater than in plain seawater [8, 13, 17, 18, 28, 32]. For comparison, the values of Eu/Sm in crusts in the region of the seamounts in the Line Islands archipelago are equal to 0.25-0.26, which is close to the upper limits of the range characteristic for the Pacific Ocean, in many regions of which the role of hydrothermal sources of REE in Mn, Fe, and other heavy metals is appreciable [11].

* * *

Seamount Krylov is a submarine volcano of the central type associated with eruption of subalkaline to alkaline, slightly differentiated basalts on the Cape Verde abyssal plain, with ferromanganese oxide-hydroxide crusts on the upper and middle parts of its rise. It is characterized by a thin-bedded to microlaminated structure; the upper 5-7 cm is mainly Fe-vernadite and Mn-feroxyhyte (Mn- δ' -FeOOH) with lesser amounts of birnessite and minor asbolane-buserite and goethite. Mn-feroxyhyte is reduced in amount toward the base of the crust.

In the basal beds ferrosilite is absent and the principal Fe oxide-hydroxide phase is goethite, while Mn oxide-hydroxides are represented by ferrihydrite. For the crust as a whole the value of Mn/Fe is relatively stable (0.64-0.89, rarely 1.13). The ferromanganese crusts are formed on a substrate composed of almost pure Fe-palygorskite, a hydrothermal alteration product of alkaline basalt hyaloclastite.

In the crusts examined, ΣREE ($\cdot 10^{-4} \%$) decreased from surface to basal bed: 2093.2 $\rightarrow$ 1453.4. There is a distinctive direct relationship ($r = 0.74$) between the lanthanide content and Mn + Fe (%), which reflects the dominance of a hydrogenic sorption mechanism in the take-up of REE by ferromanganese oxide-hydroxides.

As for the distribution of light and heavy REE in the crust section, a sharp tendency for a decrease in the light REE fraction appeared, with a corresponding increase in the heavy lanthanides, from the basal beds to the surface. It is significant that there is a high correlation ($r = 0.8$) between the values of Mn/Fe and La/Sm, i.e., the main carriers of LREE are Mn oxide-hydroxides, while correspondingly HREE are associated with Fe oxide-hydroxides: a correlation ($r = 0.89$) is found between the magnitudes of Fe and Yb/Sm. The differences established between the behaviors of light and heavy REE are interpreted to be the result of physicochemical features of the surfaces of ferromanganese hydroxides, the form of REE found in seawater, and the relative amount of hydrothermal components in the hydrothermal solution - seawater mix. LREE, which are present in seawater mainly in the form of cations, are preferentially absorbed by Mn oxide-hydroxides, which are distinguished by a low neutral charge point (pH NCP < 3), while HREE, found in seawater mainly in a complex form, are taken up mainly by Fe oxide-hydroxides, for which the pH NCP is 7.5. The decrease in the relative amount of LREE from the base to the surface of the crusts indirectly attests to the relative limitation of hydrothermal components in the seawater mix during the late stages of deposition of ferromanganese oxide-hydroxides.

The concentration of Ce and the magnitudes of Ce/Ce*, Ce/La, Ce/Yb, and Ce/Sm characteristically decrease from the basal beds to the surface. A clear correlation is seen between the magnitudes of Fe and Ce/Sm, which associates Ce with HREE. On the whole, with hydrogenic factors being of foremost importance (Ce/Ce* = 4.15-2.33), the subsidence of the seamount Krylov and the exposure of parts of the crustal deposits to interaction with seawater at great depths, where Ce is depleted, play a significant role in the variation in Ce content. This took place with movement of the Cape Verde plain away from the axial zone of the Mid-Atlantic ridge toward the African continent. It has been established [10-12] that shallow waters of the Atlantic are enriched in Ce compared to abyssal waters, which is reflected in ferromanganese oxide-hydroxide crusts. The behavior of Eu is characterized by a decrease in the value of Eu/Sm from the lower part of the crust (0.30-0.32) to the surface (0.26). When it is considered that the magnitude of Eu/Sm in hydrothermal Mn deposits of the Galapagos rift zone is 0.31 [22], and for seawater 0.26, and realizing that the composition of hydrothermal sources in the active zones of the ocean [21] are distinguished by a 1000-fold enrichment in REE with Eu/Sm approximately equal to 2.6, then the role of hydrothermal components in the seawater mix during the early stages of formation of ferromanganese crusts may be seen to be quite important.

Thus the mineralogical composition and distribution of REE in ferromanganese oxide-hydroxide crusts on seamount Krylov in the Eastern Atlantic suggest that during the early stages of formation, when this prominence was located relatively close to the axis of the Mid-Atlantic ridge, and east-west (transform) faults were active, the role of hydrothermal components was appreciable. During later stages, as the Cape Verde abyssal plain moved eastward, hydrogenic factors became more important.

LITERATURE CITED

1. I. M. Varentsov, Geochemistry of Transition Metals during Formation of Ferromanganese Ores in Modern Basins [in Russian], 25th Int. Geol. Cong., Rept. of Sov. Geologists, Mineral Deposits, Nauka, Moscow (1976), pp. 79-96.
2. I. M. Varentsov, V. A. Drits, A. I. Gorshkov, et al., Processes of Formation of Ferromanganese Crusts in the Atlantic: Mineralogy, Geochemistry of Principal and Dispersed Elements. Seamount Krylov, in: Origin of Sediments and Fundamental Problems in Lithology [in Russian], Nauka, Moscow (1989), pp. 58-78.
3. B. P. Zolotarev, Formations of the Second Layer of the Oceanic Crust [in Russian], 27th Int. Geol. Cong., Geology of the Oceans, Nauka, Moscow (1984), pp. 136-146.

4. L. E. Shterenberg, V. A. Aleksandrova, L. V. Il'icheva, et al., "Post-sedimentation alteration of oxides in ferromanganese nodules and crusts," *Izv. Akad. Nauk SSSR, Ser. Geol.*, No. 1, 80-88 (1988).
5. A. S. Aplin, "Rare earth element geochemistry of central Pacific ferromanganese encrustations," *Earth Planet. Sci. Lett.*, 71, No. 1, 13-22 (1984).
6. L. S. Balistieri and J. W. Murray, "The surface chemistry of δ -MnO₂ in major ion seawater," *Geochim. Cosmochim. Acta*, 46, 1041-1052 (1982).
7. L. S. Balistieri and J. W. Murray, "The absorption of Cu, Pb, Zn, and Cd on goethite from major ion seawater," *Geochim. Cosmochim. Acta*, 46, 1253-1265 (1982).
8. T. S. Bowers, K. L. von Damm, and L. M. Edmond, "Chemical evolution of mid-ocean hot springs," *Geochim. Cosmochim. Acta*, 49, 2239-2252 (1985).
9. H. Craig, "A scavenging model for trace elements in the deep sea," *Earth Planet. Sci. Lett.*, 23, 149-159 (1974).
10. H. J. W. DeBaar, M. P. Bacon, and P. G. Brewer, "Rare earth distributions with positive Ce anomaly in the western North Atlantic Ocean," *Nature*, 301, 324-327 (1983).
11. H. J. W. DeBaar, M. P. Bacon, P. G. Brewer, and K. W. Bruland, "Rare earth elements in the Pacific and Atlantic Oceans," *Geochim. Cosmochim. Acta*, 49, No. 9, 1943-1959 (1985).
12. H. J. W. DeBaar, P. G. Brewer, and M. P. Bacon, "Anomalies in rare-earth distributions in seawater: Gd and Tb," *Geochim. Cosmochim. Acta*, 49, No. 9, 1961-1969 (1985).
13. J. M. Edmond, "The chemistry of the 350°C hot springs at 21°N on the East Pacific Rise," *EOS*, 61, 992 (1980).
14. H. Elderfield and M. J. Greaves, "Negative cerium anomalies in the rare earth element patterns of oceanic ferromanganese nodules," *Earth Planet. Sci. Lett.*, 55, 163-170 (1981).
15. H. Elderfield, C. J. Hawesworth, M. J. Greaves, and S. E. Calvert, "Rare-earth elements geochemistry of oceanic ferromanganese nodules and associated sediments," *Geochim. Cosmochim. Acta*, 45, 513-528 (1981).
16. M. A. Haskin and L. A. Haskin, "Rare earth in European shales: a redetermination," *Science*, 154, No. 3748, 507-509 (1966).
17. G. P. Klinkhammer, P. Rona, M. Greaves, and H. Elderfield, "Hydrothermal manganese plumes in the Mid-Atlantic ridge rift valley," *Nature*, 314, No. 6013, 727-731 (1985).
18. C. Lalou, G. Thompson, P. A. Rona, et al., "Chronology of selected hydrothermal Mn oxide deposits from the transatlantic geotraverse 'TAG' area, Mid-Atlantic ridge 26°N," *Geochim. Cosmochim. Acta*, 50, 1737-1743 (1986).
19. Y. H. Li, "Ultimate removal mechanisms of elements from the ocean," *Geochim. Cosmochim. Acta*, 45, 1659-1664 (1981).
20. V. E. McKelvey, N. A. Wright, and R. W. Bowen, *Analysis of the World Distribution of Metal-Rich Subsea Manganese Nodules*, U. S. Geol. Survey Circ. 886 (1983).
21. A. Michard, F. Albarede, G. Michard, et al., "Rare earth elements and uranium in high-temperature solutions from East Pacific Rise hydrothermal vent field (13°N)," *Nature*, 303, No. 5920, 795-797 (1983).
22. A. A. Migdisov, B. P. Gradusov, N. V. Bredanova, et al., "Major and minor elements in hydrothermal and pelagic sediments of the Galapagos mounds area. Leg 70," *Init. Repts. DSDP, Wash.*, 70, pp. 277-295 (1983).
23. M. R. Palmer, "Rare-earth elements in foraminifera," *Earth and Planet. Sci. Lett.*, 73, 285-298 (1985).
24. M. R. Palmer and H. Elderfield, "Rare earth elements and neodymium isotopes in ferromanganese oxide coatings of Cenozoic foraminifera from the Atlantic Ocean," *Geochim. Cosmochim. Acta*, 50, No. 3, 409-471 (1986).
25. D. Z. Piper, "Rare earth elements in ferromanganese nodules and other marine phases," *Geochim. Cosmochim. Acta*, 38, 1007-1022 (1974).
26. D. Z. Piper and P. A. Graef, "Gold rare earth elements in sediments from the East Pacific Rise," *Geochim. Cosmochim. Acta*, 17, 287-297 (1974).
27. D. E. Ruhlin and R. M. Owen, "The rare earth element geochemistry of hydrothermal sediments from East Pacific Rise: examination of a seawater scavenging mechanism," *Geochim. Cosmochim. Acta*, 50, No. 3, 393-400 (1986).
28. V. Tannicliffe, M. Botros, M. E. de Burgh, et al., *Hydrothermal Vents of Explorer Ridge, Northeast Pacific, Deep Sea Res. Pt A*, 33(3A) (1986), pp. 401-412.
29. D. R. Turner, M. Whitfield, and A. G. Dickson, "The equilibrium speciation of dissolved components in freshwater and seawater at 25°C and 1 atm. pressure," *Geochim. Cosmochim. Acta*, 45, 855-881 (1981).
30. I. M. Varentsov, N. V. Bakova, Yu. P. Dikov, et al., "Synthesis of Mn, Fe, Ni, Co oxide-hydroxide phases on manganese oxides: on the model for transition metal ore formation in Recent basins," *Acta Mineralogica-Petrographica, Szeged.*, 24, No. 1, 63-90 (1979).

31. I. M. Vrentsov, B. A. Sakharov, V. A. Drits, et al., "Hydrothermal deposits of the Galapagos rift zone, Leg 70: mineralogy and geochemistry of major components," Init. Rept. DSDP, Wash., 70, 235-268 (1983).
32. K. L. von Damm and J. M. Edmond, "Chemistry of submarine solutions at Guaymas Basin, Gulf of California," *Geochim. Cosmochim. Acta*, 49, 2221-2237 (1986).
33. M. Whitfield and D. R. Turner, "Ultimate removal mechanisms of elements from the ocean - a comment," *Geochim. Cosmochim. Acta*, 46, 1989-1992 (1982).

SEDIMENT GENESIS IN AREAS OF HOLOCENE PEAT ACCUMULATION, KHOLKHIDSK (COLCHIS) LOWLAND

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UDC 552.577(479.224)

Detailed lithofacies analysis was employed in establishing the makeup of Holocene sedimentary sequences of the Colchis Lowland. An incomplete New Black Sea sedimentary cycle of sediment and peat accumulation has been established involving facies parageneses and depositional zones. Alternation of these zones in a westerly direction is accompanied by composition changes in the peat bed from complex to simple makeup, with a thickening toward the center of the peat swamp, as well as changes in plant components of the lignitic peat group. Peat deposition was preceded by deposition of continental deposits.

The Colchis Lowland is a unique, clearcut model of a Holocene coastal plain with sediment and peat accumulation. Structurally, it is associated with inherited compensation of the Rioni intermontane basin [6]. Topographically, the area consists of three zones: a hilly piedmont upland, the Colchis Lowland that stretches to the Black Sea from which it is separated by a narrow sandy beach zone. Tectonically, there are three subbasins in the Rioni Intermontane Basins: the Potiisk Basin between the South Megrel'sk Ridge on the north and the Gruisk Ridge on the south; to the west, it is bordered by the shoreline between the Rioni and Supsa Rivers. The Abkhaz-Mergel'sk and Kobuletsk Basins occur north, respectively south, of the Potiisk Basin. In each of the basins the Colchis Lowland is the site of peat accumulation. The Lowland stretches along the Black Sea for a distance of 180 km between the Kodori River and Cape Tsykhisdzryi, south of the Kintrishi River. The Lowland is ≈ 60 km wide in the central part of the Rioni intermontane basin (Potiisk Subbasin) from the shore, in an inland direction, toward the mouth of the Tskhenis-Tskali River, or the town of Samtredia. It narrows sharply toward the south (Kobuletsk Plain) and especially in a northerly direction (Abkhaz-Mergel'sk Plain). Its shape, therefore, is triangular. The base of the recent peat beds beneath the Lowland lies below the level of the Black Sea. This indicates subsidence, confirmed also by a whole set of buried, thin high-ash peat horizons in the Quaternary section at depths between 9-120 m. The Colchis Lowland has a subtropical climate.

An extensive literature is available on the investigated Lowland deposits, including those of the Rioni intermontane basin. Authors include Vakhaniya [2], Gamkrelidze [3], Dzhanelidze [4], Dumitrashko [6], Laliev [9], Mamaladze [11], Makedonov [10], Milanovskii [13], Motsereliya [14], Tsereteli [19] and others. These studies dealt with many generally or partially discussed overall geological problems, associated with tectonics, stratigraphy and Mesozoic-Cenozoic geological evolution of the region, based on a general paleogeographic reconstruction. They have also displayed the absence of available detailed lithofacies definition of deposits and facies changes, associated with Holocene sediment and peat accumulation during formation of the Colchis Lowland. The paleogeographic concepts as applied to this period have also been incomplete. Inaccuracies were detected and inconsistencies between

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the stratigraphic charts of various authors were revealed. Their evaluation was difficult due to the lack of geological correlations of sections and drillhole logs. Incorrect statements have been made on the depositional facies of sediment units that were associated with peat formation. Consequently, our task was the identification of categories of sediment genesis and the facies conditions of sediment and peat formation. The aim included learning about their parageneses and areal distribution patterns within the context of Holocene paleogeographic evolution of the Colchis Lowland. The Lowland was considered a model of the early coal development stage in an intermontane depression that was open to the seashore.

Work Performed and Study Methods. We have undertaken a detailed study of New Black Sea Stage deposits in the Colchis Lowland. In our view, these deposits represented the last sedimentary cycle, associated with the thickest peat development. More than 600 drillholes were studied to depths of 10-25 m. A small number of holes were 250-300 m deep. For a small number of drillholes descriptions, columnar sections, cross sections have been utilized from reports and dissertations of Tvalchelidze [17]. The density of drillhole control (the holes were several tens of meters deep in areas of peat deposits and hundreds of meters deep outside of them) allowed compilation of lithofacies profiles with acceptable detail, both parallel (29 sections) and perpendicular (12 sections) with the Black Sea coast (Fig. 1). Radiocarbon dates of peats provided additional control for a very accurate correlation of deposits within the cross section, despite the highly variable depositional facies of the sediments involved.

The sediments were subjected to lithofacies analysis, as developed for consolidated Paleozoic and Mesozoic coal-bearing deposits [7, 8, 18]. However, it was difficult, occasionally impossible, to obtain from the drill samples even such basic diagnostic data as textural features for genetic sediment classification in unconsolidated sands and saturated, layered aleurolites and clays. In cores, analogs of such sediments have been well consolidated. For this reason, it became necessary to search for additional diagnostic lithofacies features that would allow better demarcation and characterization of genetic types and facies of sediments, still unaffected by lithification. Naturally, this and the lithofacies method can not be treated dogmatically. This may facilitate the approach toward given, specific research objectives.

The most effective method involved microscopic study of sediments in transparent thin sections, polished on both sides. A series of reliable genetic sediment features have thus been established: plant content, degree of consolidation, size of vegetative components, degree of preservation of their tissue structures, morphology of disintegrating tissues, presence of water plant remains, fungus filaments and sclerotia, quantity and composition of the diatom flora, microtexture characteristics, quantity and composition of authigenic and heavy minerals. Additional diagnostic features included the residual chlorine composition in the sediments, along with the above-cited features that had been microscopically identified.

Lithofacies Sediment Characteristics

Detailed lithofacies analysis, conducted for the first time on the Holocene Colchis Lowland deposits, identified ten lithological sediment categories, representing 59 genetic types within 27 facies. Within the peats, 27 genetic categories and nine facies have been identified. The selection principles of the cited taxonomic units have repeatedly been discussed in the literature, in particular since the 1940s [7, 8], although they were defined more completely by Timofeev [18] only later. We briefly characterize the sediment and peat-forming facies (a more detailed description of the lithological and genetic sediment types is to be given in a monograph, under preparation). Eight genetic sediment groups were established, four of which apply to detrital sediments, four to peats.

Alluvial Deposits (A). These are the most widespread in the Colchis Lowland. Seven depositional facies were identified, distinguished by genetic characteristics. Facies ARG is characterized by mountain stream channel deposits and their alluvial fans that include gravelly-bouldery alluvium types, with admixture of coarse grained, very poorly sorted sands of angular grains. Facies ARP includes sediments near river mouths and adjacent Colchis coastal plain stream channels that between themselves displayed differences in granulometric composition. The first type usually is more coarse-grained: it contains medium and coarse-grained sands, occasionally gravelly, with admixtures of coarse gravel. The alluvium of the second facies is represented mostly by fine-grained sands or coarse aleurolites. Layering was impossible to establish in the loose drill samples. It was noted only in outcrops where

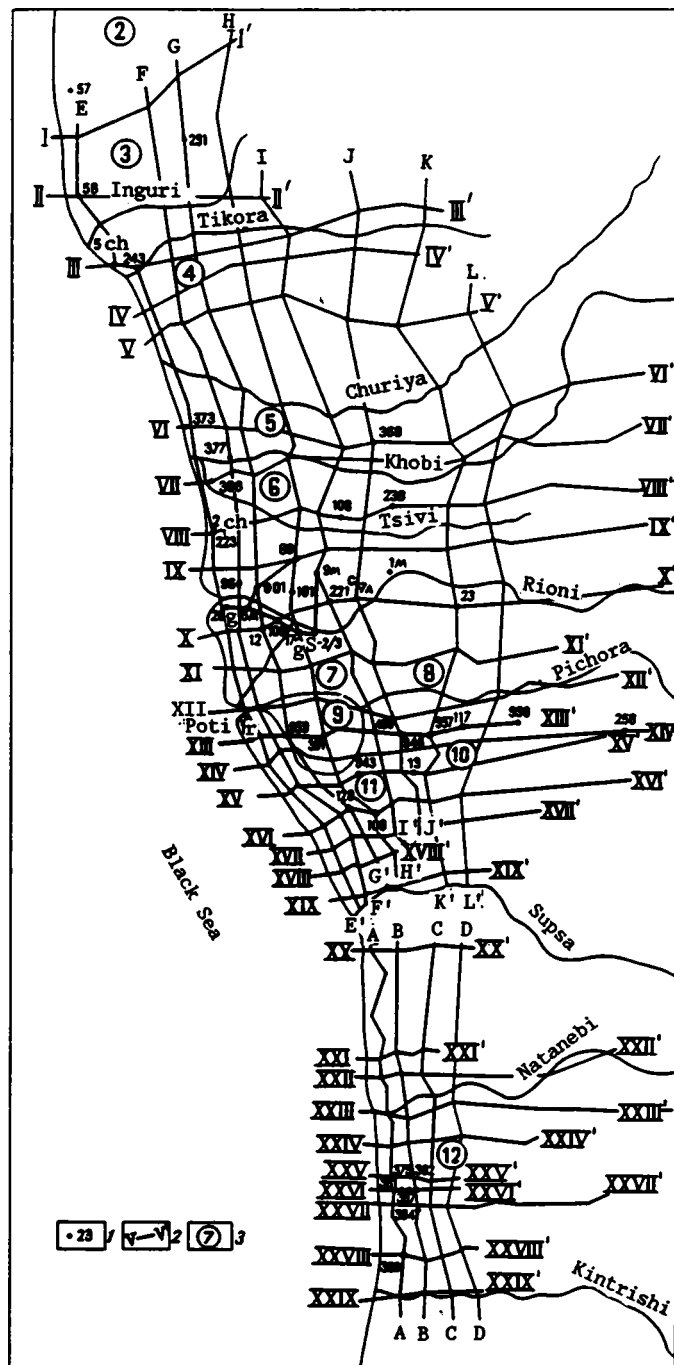


Fig. 1. Location map of drillholes and cross sections in the Colchis Lowland peat deposit area: 1) drillhole number; 2) cross section line; 3) peat deposits; circled number: 2 - Gagidsk; 3 - Zorgatsk; 4 - Anakliisk; 5 - Churiisk; 6 - Nabadask; 7 - Poti; 8 - Pichora; 9 - Paliastomsk; 10 - Imnatsk; 11 - Maltakvsk; 12 - Kobuletsk.

it occurred in the form of various large- and small-scale cross stratification types [1]. The ARP facies is also well defined by its poorly sorted sand fraction, with poor rounding, the presence of minor amounts of large plant stalks and stems, the absence of sapropelic matter, fungi, faunal remains, authigenic minerals and minor concentrations of heavy minerals.

Sediments of near-stream channel facies (APP) and the interior floodplain facies (APV) possess many common features, such as, the poor sorting with an average rounding index of the sand grains, grayish-light blue, often grayish-dark blue color, most importantly, horizontal layering, caused by alternation of thin laminae of different granulometric composition,

distinctly visible even in the plastic drillcores. However, despite the similarity of most genetic features, sediments of the two facies differ sharply in granulometric makeup and characteristics of their minor plant content. The APV facies contain more clay and are characterized by tabular forms of the mostly fine-grained attrital coal matter.

Sediments of forested (APZ) and open (APO) lake areas in the floodplain differ in all genetic characteristics from sediments of this group's other facies, while displaying similarities between themselves. They are well sorted, contain much clay, display poorly- to well-expressed horizontal layering and contain many sponge spicules and remnants of fungi that are instrumental in the conversion of the well preserved humitic plant matter. The basic difference is in the composition of plant matter: woody in sediments of the APZ facies, while grassy in facies APO, with admixtures of aqueous plant tissues. These deposits often occur at the top of peat layers. In such instances, the leaf- and stem-matter is particularly abundant; the sharp layering reminds one of layer cake. Despite differences between their genetic indicators, facies APZ and APO of the alluvial group paragenetically are closely related in floodplain settings.

Oxbow lake (APS) sediment features in terms of granulometry, occasionally of sorting, and with regard to layering are similar to those of the interior floodplains, floodplain lakes. In their significant fungus and organic matter content they resemble floodplain lakes, but the lakes' variable attrital content makes the lake sediments similar to interior floodplain facies. Sediments of the APS facies also have unique diagnostic features: strongly decomposed plant matter, associated humo- and saprocollinite, and minor amounts of diatoms.

Group of Proluvial Deposits (P). This is piedmont alluvial fan and surface sediment train (slopewash) facies (PPK) that characterize intermittent stream. Their diagnostic features have been described in the literature [18].

Group of Lake Deposits (O). These are associated with peat deposits. Sediments of facies OSP, OSS, OSO of this group are characterized by sapropelic lake sedimentation on the coastal plain that preceded peat accumulation. They occur locally and only in the Kobuletsk Basin. Some of the sediments have higher clay, others higher slit content, rarely well-sorted sandy composition, usually accounting for the nonlayered sediment structure. They differ in a number of genetic features, such as greenish-gray coloration, the abundance of fresh water diatoms that often leads to transition toward diatomites, mixed with helinitic and saprocollinitic organic matter that is quite abundant in the sediments. Despite similarities between most genetic characteristics, sediments of each of the three facies possess singular diagnostic characteristics that had been used in their identification. Sediments of facies OSP are characterized by only freshwater diatoms, while sediments of facies OSS and OSO contain also salt-water forms in different proportions. Brackish-marine diatoms also occurred. Differences between sediment types, based on diatoms of various biotype settings, indicate differences between salinities of the different lake basins. Salinities increased from facies OSP to facies OSS and OSO.

Sediments of facies OZD and OZT were defined by genetic characteristics, in certain instances by increase of lake basin areas, recalling soil formations. They include remnants of root systems, occasionally microaggregates. In other instances, muddy woody or grassy swamp deposits occurred in the top peat interval. Sediments of both facies include much organic matter, mostly with humus content. They are characterized by high clay content, good sorting, usually uneven layering, significant amounts of fungus remains that are combined with considerable concentrations of gelinitic-collinitic matter in the humus fraction. The type of floral matter is characteristic of given sediment types: in facies OZD it is mostly woody; in facies OZT, mostly grassy. Tissue of water plants in sediments of both facies indicates their deposition in lakes. Therefore, they are closely related to sapropelic lake deposits, although most of their genetic indicators are quite varied.

Group of Continental Coastal Plain Swamp Deposits (Soil Formations-P). This group includes sediments of five swamp soil and subsoil facies. Each is distinguished by specific genetic conditions of the substrate and the floral composition of swamp phytocoenosis. Thus, soil formations of facies PDD and PDT developed in river valleys in swampy, respectively, forest and meadow settings. Soils of facies PSD formed through reworking of woody plant vegetation of sapropelic lakes, while soils of facies PMD and PMT in marine nearshore settings. Woody and grassy phytocoenoses played the main role in the conversion of this setting in swampy facies. Deposits of this facies group are transitional toward peaty lake deposits. All these facies of the soil formations are characterized by mottled colorations, due to the

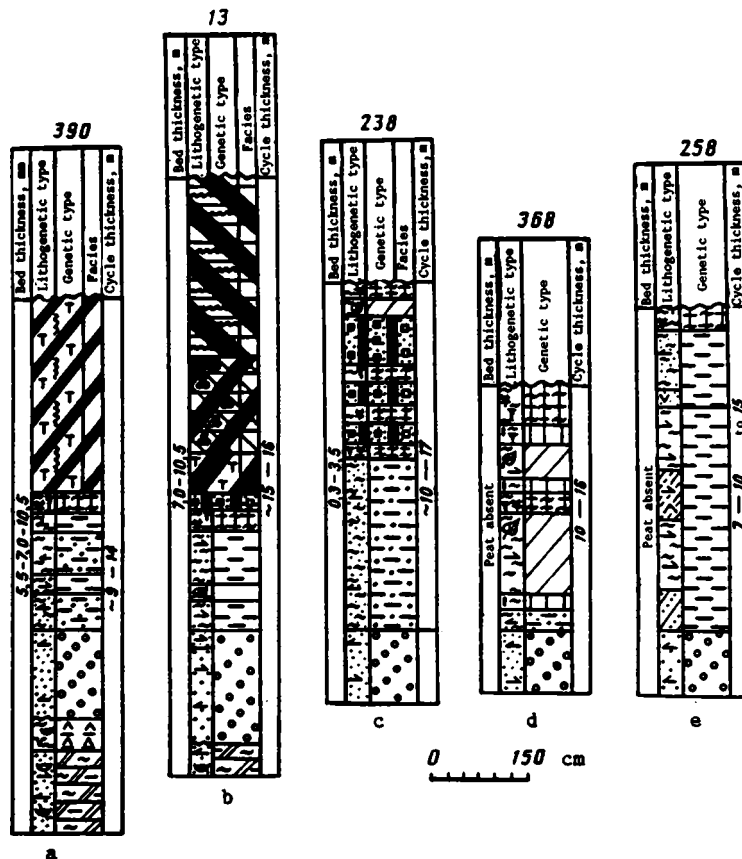


Fig. 2. Facies types of cycles: a) alluvial-swampy plant fragments-peaty; b) alluvial-swampy plant fragments-mosses; c) alluvial-swampy, woody-peaty; d) alluvial-swampy; e) alluvial. Conventional symbols: Fig. 4.

alternation of gray, beige, rusty-brown colors, lumpy, often microaggregate-type structures, usually significant organic matter content with admixtures of significant root tissue matter, fungus remains, an iron hydroxide enrichment, and the presence of colomorphic clays that upon contact with roots became bleached.

The group of marine nearshore deposits (M) includes low-energy bay and lagoon areas (MPZ and MPL) and their higher-energy shallow water portions (MPR and MPP). This also includes nearshore, higher energy shallow marine areas (MMV, MMK, MMM and MMD) and distal portions of the marine nearshore zones (MUM) that characterize inter-delta and delta margin conditions.

More of the genetic indicators in sediments of facies MPZ and MPL are similar than different. They differ in the clay content, vs. the better sorted aleurolite content. Grayish-brown, brown chocolate-brown hues, with characteristic olive green coloration are most typical. Large amounts of mixed organic matter also occurs with significant amounts of gelinitic and saprocollinite. Intensive gelitanization of the attritus and tissue fragments, the presence of in-situ pyrite-enrusted marine micro- and macrofauna were also typical. The unique character of the sediments of both facies resulted from the quantity and ecotypes of enclosed diatom algae. Only minor amounts of mostly brackish-marine algae occurred in facies MPZ, while sediments of facies MPL contained an abundance of diatoms, representing the same biotopes.

Sediments of facies MPR and MPP of this group in the Colchis Lowland occur only occasionally. Essentially, they contain moderately sorted sandy matter with low grain rounding values. Due to the loose consistency and viscous state of the drill samples, layering could almost never be determined in fresh samples but when dried, lenticular, low-angle layering, and cross lamination (sediments of facies MPR), and horizontal, horizontal-wavy and trough cross-stratification types (in sediment of facies MPP) have been observed. Generally, both facies contain large quantities of organic matter. Total absence of gelinitic-collinite

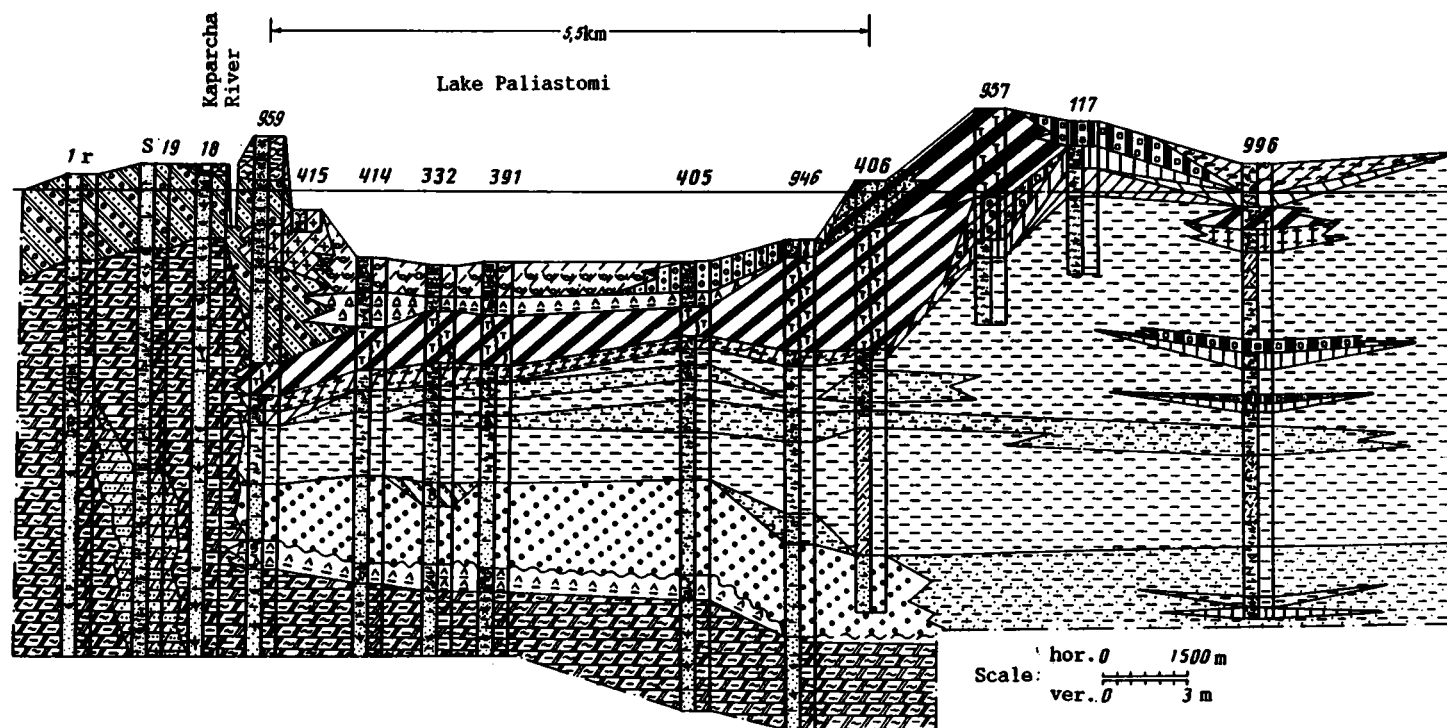


Fig. 3. Schematic cross section of facies. Poti, Pichora, and Imnatsk peat deposits of the Rioni intermontane basin. Line XIII-XIII'. Symbols: Fig. 4.

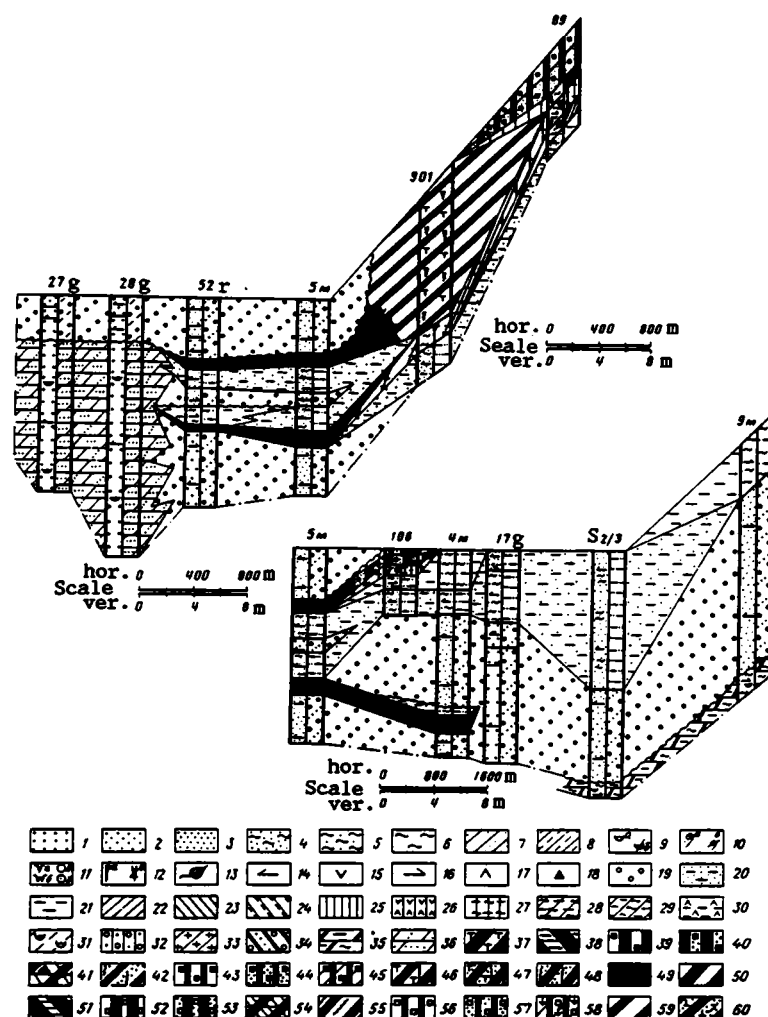


Fig. 4. Composition of peat beds of the New Black sea cycle in the Rioni River channel migration zone. Lithological sediment types: 1-3) coarse, intermediate and fine-grained; 4-5) aleurolite of coarse (sandy loam) and fine-grained (loam); 6) clay; 7) interlayered aleurolites and sands; 8) interlayered aleurolite and clays. Inclusions: 9) colluscan shell fragments (a - complete; b - fragmental); 10) foraminifer tests (a - complete; b - abundant) and fresh-water (c - few; d - fragments); 11) diatoms; brackish (a - few; b - abundant); 12) soil microaggregates (a) and rhizome remnants (b); 13-15) leaf, wood and grass tissue fragments, respectively; 16-17) durain: woody and grassy, respectively; 18) iron hydroxides. Depositional facies: 19) gravelly-sandy sediments along river mouths and adjacent channels of floodplain streams (lower river reaches)-ARP; 20) sandy-silty floodplain sediments near river mouths-APP; 21) sandy-silty-clayey sediments of interior floodplain areas (APO); 22) silty-clayey lake sediment in forested floodplain zones-APZ; 23) open lake sediments in grassy floodplain zones-APO; 24) silty-clayey oxbow lake sediments-APS; 25) silty-clayey sediments in lakes, overgrown by vegetation, and in woody-peat-filled swamps-OZD; 26) silty-clayey sediments in lakes, overgrown by vegetation and in grassy-peat-filled swamps-OZT; 27) silty-clayey sediments of forested, swampy river valley areas; 28) clayey sediments of swampy river valley grasslands-PDT; 29) sandy deposits of swampy seashore grasslands-PMT; 30) silty-clayey sediments of stagnant coastal bay areas-MPZ; 31) silty-clayey deposits of

matter resulted from the absence of fungus remnants and the low degree of decomposition of such plant substances as attritus in facies MPR sediments that are fragmental, as are the tissue fragments that predominate in sediments of facies MPP. The small number of brackish diatoms in contrast with abundant faunas, provides a sharp distinction between sediments of facies MPP and MPR.

High-energy nearshore marine deposits are the most widespread among the nearshore marine deposits of the Colchis Basin. These include beach and coastal ridge deposits (facies MMV), underwater shoals and bars (facies MMK), rippled shallow water facies (MMM) and underwater delta facies (MMD). The characteristic genetic indicators of all these facies include a predominantly sandy composition, the most frequently good, less commonly moderate only occasionally poor sorting of the sediments, and a high rounding index of the sand grains, and the presence of marine faunas. They also incorporated humus-type organic matter and contained no diatoms, fungus remains and chlorine residue. The type of layering was vital genetic in each of the discussed facies. Layering occurred in sediment outcrops and in peat pits. They could reliably be also defined for drill samples in which, due to the mechanical impact on the unlithified, pliable sediment matter, layering became less distinct. Additional, individual genetic features may be represented by differences in the organic matter content, its floral composition, the presence of marine facies elements, pyrite, residual chlorine, and of heavy minerals.

Sediments of facies MUM were more unique in terms of their genetic features than sediments of the rest of this group. They consist of well-sorted clays and minor amounts of mixed organic matter. The humus components were represented mostly by attritus; the sapropelic component by saprocollinite. Sediments of this facies differ in the abundance of the marine faunas (shell fragments), along with coccoliths and brackish-marine diatoms. The sediments were enriched in pyrite and authigenic calcite. They occurred only in one of the studied drillholes in which they displayed very clearcut genetic characteristics.

Peats of forested (alder-bearing) peat swamps and grassy peat swamps were the most widespread among the peat swamps facies of the Colchis Lowland. Moss peat bogs played a very minor role. An intermediate role was placed by forested (alder-bearing) peat swamps. Each genetic peat category may be subdivided on the basis of phytocenosis and thanatocenosis,

stagnant lagoonal seashore areas-MPL; 32) silty-sandy sediments of small river deltas in bays and lagoons in the marine coastal zone-MPR; 33) sandy bay deposits in areas adjacent to the seashore-MPP; 34) gravelly-sandy marine beach and shore embankment deposits-MMV; 35) sandy, surf zone deposits, shallow, flat nearshore bottoms-MMM; 36) sandy marine delta deposits-MMD; Genetic peat types: 37) rihozome-sedge gelinite-telinite; 38) Sphagnum imbricatum-gelinite-telinite peat; 39) helinite-posttelinitic alder peat; 40) same, with ash-content; 41) molinitic sphagnum peat, gelinitic-post-telinitic; 42) reed-helinitic-posttelinitic, ash-bearing; 43) alder peat, gelinitic-precollinitic stage; 44) same, ash-bearing; 45) alder-sedge peat, gelinitic-precollinitic; 46) sedge-reed gelinitic-precollinitic; 47) same, ash-bearing; 48) reed-gelinitic-precollinitic, ash-bearing; 49) peat. Peat-forming facies: 50) grassy, gelinitic peats of strongly saturated lowlands peat marshes; 51) mossy gelinitic-telinitic peat of transitional upland peat swamps; 52-53) gelinitic-posttelinitic alder-bearing, weakly flooded, stagnant, and flooded-circulating, intensively drained lowland peat swamps; 54) grassy-mossy, gelinitic-posttelinitic peats of strongly flooded, stagnant transitional peat swamps; 55) highly flooded, grassy, circulating, lowland peat swamps; 56-57) alder-bearing, helinitic-precollinitic peats of slightly flooded, stagnant and flooded-circulating strongly drained lowland peat swamps; 58) grassy-alder-bearing, gelinitic-precollinitic peats of medium-flooded, stagnant, periodically drained lowland peat swamps; 59-60) grassy, gelinitic-precollinitic peats of strongly flooded, stagnant, and submerged-flooded and circulating lowland peat swamps. Width of band in symbols 41, 42, 54, 55 corresponds to posttelinitic structure in grassy peats, already telinitic in composition (symbol 37), broader than for precollinitic types (symbols 46-48, 59, 60).

characterized by a combination of well-defined peat-accumulating facies, corresponding to genetic peat categories. Distinction is based on the degree of decomposition of the lignitic and cellulose-bearing tissues, on the peat (telinitic, post-telinitic, precollinitic and collinitic types) and mineral components. A more detailed designation of the individual taxonomic units in the peats and their genetic classification will follow in a separate paper.

Lithofacies Characteristics of Peat Deposit Sections of the Colchis Lowland and Their Composition Patterns. The numerous profiles (Fig. 1) portrayed the lithological and facies composition changes that occurred in horizontal and vertical directions in the Holocene deposits in the Colchis Lowland subsurface. The profiles extend from the town of Samtrediya for a distance of ≈ 50 -60 km and along the Black Sea shore from the town of Ochamchir (north) to the town of Kobuleti (on the south). The Holocene Colchis Lowland peat deposits contain Old Black Sea Stage fine- and medium-grained sands, deposited in rippled shallow marine bottom facies (MMM) with which coarser, higher-energy shallow marine deposits are also associated (MMK-underwater bars, ridges, and shoals). Sands and aleurolites of marine coastal deltas (MMD) and certain less frequent (MUM) facies also occurred. Most of these sediments contain the following fauna elements of the Old Black Sea Stage at various Colchis Lowland locations: Corbula mediterranea, Chione gallina, Hydrobia ventosa, and Cardium edule (identifications by P. V. Fedorov). Locally, the Old Black Sea Stage deposits are topped by silty-clayey lenses of bay-lagoonal deposits (facies MPZ and MPL), occasionally by indications of swamp development, indicative of increasing influence by the land.

The Old Black Sea Stage deposits upward alternate with sediments of the single cycle of the peat- and sediment-forming, incomplete New Black Sea Stage. Its incomplete character is shown by the fact that it includes only the transgressive hemicycle and the regressive sequence is absent above the peat unit. This was due to the overall paleogeography of the peat-accumulating Colchis Lowland area and related to the continuing Black Sea transgression during the early Holocene. Sections at the base of the New Black Sea peat and sediment cycle almost always contain quite thick (4-10 m, occasionally thicker) river valley alluvial channel (facies ARP) or mountain river channel deposits (facies ARG, Kobuleti). The Black Sea coastal strip is an exception. This alluvium overlies an erosional unconformity, cut into various horizons of upper Old Black Sea Stage deposits. The radiocarbon age of bay-lagoonal deposits that accumulated during surface erosion and occur at 19 m depth in Drill-hole #23 (right bank of the Rioni River, 13 km from the Black Sea shore) was 7900 ± 60 yr. B.P. (Tb-86). Consequently, the channel alluvium at the base of the New Black Sea Stage should be somewhat younger. This means that in this part of the Colchis Lowland the new Black Sea Stage occurred slightly later, sometime 7500-7700 years ago. This may be approximately true also of the age of all sediment and peat units of the New Black Sea cycle.

In terms of the section position in the overall context of the structural and paleogeographic configuration of the Colchis Lowland, the New Black Sea peat and sediment accumulation cycle has a varied makeup. It is characterized by an uneven collection of genetic and facies types, with different parageneses, number of layers, their thickness and varying thickness values of sediments in the cycle. Sediments of various facies, classified in the preceding, were combined in the sequence of the New Black Sea cycle. These form a succession, composed of eight clearly defined, incomplete facies types of the cycle, with well-defined composition and sequence order of the units. The sequence consists of alluvial (Fig. 2e), alluvial-swampy (Fig. 2d), alluvial-woody-swampy (Fig. 2c, Fig. 3, drillhole #117), alluvial-swampy, grassy-mossy peaty (Fig. 2b), alluvial-lacustrine-moss peaty (boggy), alluvial-swampy, peaty marsh (Fig. 2a), alluvial-swampy-lagoonal peaty marsh (Fig. 3, drillhole #414-405), alluvial-swampy-nearshore-marine peaty marsh (Fig. 3, drillhole #959). The sub-units of the sediment cycle, their terminology, definition and composition has been widely discussed [7, 8]. However, we are basing our findings on work by Timofeev who in his fundamental monograph on the Jurassic coal-bearing formations of southern Siberia [18] and subsequent works developed an exhaustive system.

In tracing components of the cycle in the Colchis Lowland cross sections, the pattern of changes became obvious from the east toward the west. Sedimentation zones were thus established in the Colchis Lowland. The sequence in each zone is characterized by a given portion of the described categories of the cycle. The zones are the following: rear margin of the Colchis Lowland (the zone that surrounds the peaty area), the peat zone, peat swamp zone, central portion of peat swamps, and the shore zone of peat accumulation. In addition, the zones of maximum subsidence have been established along the southern and northern margins of areas of sediment deposition in the central peat swamp area of the Potiisk Basin. Deposits

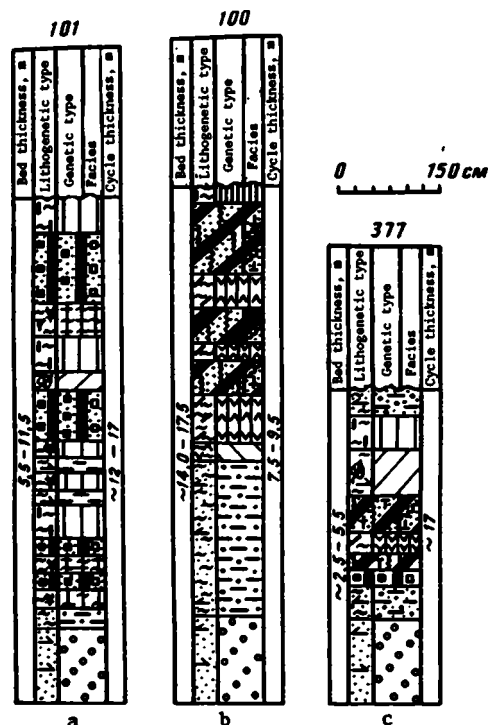


Fig. 5. Sections containing the New Black Sea cycle along the Khobi and Tsiva river channels.

of the new Black Sea cycle have been identified within these sediments on the basis of their facies characteristics. Areas of maximum subsidence has also been located in the Kobuletsk Basin of the Colchis Lowland. They occur in the center and along the southern and northern basin margins and display differences in the sediment patterns.

The cross sections indicated that the essential peat units, in terms of their greatest thickness and largest area occupied, in the Tioni interfluvial basin occurred in the Potisk Basin, within the most deeply buried areas. For this reason, the investigation of lithofacies composition of the sections and their composition patterns commenced in this area and proceeded from south to north toward the basin margins, in the direction of the Abkhaz-Mergel'sk and Kobuletsk Basins.

The Potisk Basin occurs in the central portion of the Rioni interfluvial depression and of the Colchis Lowland. Its center occurs approximately at the main Rioni River channel (Fig. 1). The channel nearly overlaps with the axis of the depression. The modern Lake Paliostomi occurs in the nearshore area, south of Rioni River. The lake is remnant of a bay of the New Black Sea basin. Additional larger rivers and a number of smaller streams cross the Potisk Basin, emptying into the Black Sea. These include the Tikora, Inguri, Churiya, Khobi, Pichora, Maltakba and Supsa Rivers that also serve as borders of major peat swamps with commercial peat deposits. These from north to south are the Tikora, Anakliisk, Pichora, Paliastomsk, Imnatsk and Maltakvsk deposits.

The Holocene sequence in the area around the Colchis Lowland in the Rioni Intermontane depression of the Potisk Basin consists essentially of alluvial deposits with proluvial-deluvial and fluvioglacial deposits. These generally are 10-15 m thick [19]. They have been studied only in cursory fashion and their characteristics have not yet been established. At the base of the section the bouldery-gravelly, pebbly and bouldery-angular gravelly clayey sediments toward the west, southwest gradually become depleted in gravelly, sandy and aleuritic matter. As the bouldery-gravelly beds became thinner, the thickness of the other beds increased. Thus, the Holocene section eventually becomes sandy-aleurolitic, part of the whole alluvial complex that was formed by the various streams. Part of them forms the incomplete alluvial category within the New Black Sea sediment cycle (Fig. 2e). Further westward, toward the rear rim of the Colchis Lowland, toward the transitional zone around the belt of peat accumulation, the lower part of the section contains fine-grained, rarely medium-grained, poorly sorted sands (occasionally with pebbles), typical of lowland river channel alluvium

(ARP). In the quite thick (maximum 9 m) upper portion of the cycle lenticular clay layers, aleurolites, fine sands form the sequence. They and their interlayered combinations formed mostly in stream channels and in the interior floodplain (APP and APV). In the transition zone of the Colchis Lowland peat accumulation area, this alluvial member of the cycle alternates with the facies group of the incomplete alluvial-swampy facies types of the cycle that occupies almost the entire peat-forming Colchis Lowland area, with the exception of the littoral zone. The marginal zone of peat accumulation that fronts the peat swamp zone includes the section with alluvial-swampy deposits of the New Black Sea cycle (Fig. 2d). In contrast with alluvial type units of the cycle, the upper channel-alluvium deposits are complex, belong to various subfacies, and consist of repeated alternations of beds and mutual facies transitions, paragenetically related to floodplain deposits (APP and APV). Varying amounts of lake deposits (APZ) are also included, with associated portions of soil horizons (OZD), rarely accompanied by laminae of alder peat with high ash content and posttelinitic structure. Coarse- and fine-grained aleurolites and clays are the most widespread in this area. Rarely, fine-grained sands and their alternations represent various genetic types of the interior floodplain and zones that adjoin river channels. They differ in characteristic features. Usually, they overlie channel sands at the base of the cycle. Lake deposits are typical in this cycle of this area, represented essentially by clays and aleurolites of the APZ facies. Their enrichment in wood tissue remnants emphasizes the association of these sediments with forested floodplain. Sediments of facies APZ are essentially widespread in sections that occur near small river channel of the Munchini and Tsini type or near major river banks (e.g., Khobi River), where occasionally they are covered by and associated with clay deposits of facies OZD.

Further west, in the Colchis Lowland peat swamp-margin zone of sedimentation, i.e., where the peat deposits first occur, the alluvial-swamp category of the cycle is replaced by alluvial-swampy, woody peat type (Fig. 2c and Fig. 3, drillhole #117). The section starts with fine-grained river channel sand (ARP), the western continuation of the sand horizon at the base of the alluvial-swamp unit of the cycle. The middle portion of the cycle almost always consists of aleurolites with occasional fine-grained sands, representative of all genetic types of the floodplain deposits and their facies. Most frequently, the upper part of the floodplain deposits were reworked in the woody soil; less frequently they are overlain by clay deposits of facies OZD, and even more infrequently they are topped by clayey lake deposits of forested floodplains (APZ). In principle, these are not typical of the section of this category of the cycle. The sequence of floodplain deposits in the subject area consists of a sediment complex of several facies. Their base consists of a generally complex peat unit. The unit consists of 0.25–0.75 m, rarely 1–2 m thick beds that alternate at various rates. The beds consist of woody, rarely grassy peats with clay layers, mostly fine-grained aleurolites of facies OZD, PDD, sometimes of floodplains and lakes, and associated with muddy peat swamps. The discussed sediments in the polyfacies complex display facies transitions within short distances, or more frequently, simple pinchouts, forming lenses along the peat swamp margins. The primary sediments were introduced by stream floods into the peat swamp and by pluvial sheetwash from dry valleys. The thickness of the complex peat bed at the type location of this cycle type varies within a wide range of 0.3–5.5 m. The base of the thickest peat beds was radiocarbon-dated 5240 ± 60 yr B.P. (GIN-1235, Churiisk Deposit) and 6170 ± 270 yr. B.P. (GIN-1394, Nabadsk Deposit). At certain locations in the peat swamp margin area the bed is of simple composition; it consists of a single, thin unit (0.5–1.0 m; Fig. 3, drillhole #117). Toward the east, the rear margin of the peat area, the peat units usually pinch out and grade into soil and lake deposits. Rarely, they are preserved as very thin layers. Often, the peat beds in sections in this cycle type became buried under aleurolitic deposits of facies OZD, PDD or floodplain deposits, although frequently they remain exposed. Woody, gelinitic-posttelinitic rarely telinitic, frequently silty-clayey peats are typical of peat layers in the cycle. Woody gelinitic-precollinitic components, both of the ashy and the low-ashy category, also occur. However, in areas with streams, the cycle sequence displays reedy gelinitic-precollinitic peats with increased ash content (e.g., Anakliisk Deposit at the contact with waters of the Mukhurzhinka River and the Pichorsk Deposit at the peaty left bank of the Pichora River).

Sections of the New Black Sea sedimentation cycle, located in the central zone of peat swamps where the Potiisk Basin reaches its greatest depths, differ in composition and facies makeup from sediments located along their northern and southern basin margins. In those areas, they uniformly represent alluvial-swampy, reedy- or reedy-mossy peat facies units of the cycle (Fig. 2b and Fig. 3, drillholes #406 and 957). This facies group can be easily

traced from the Churiisk to the Imnatsk Deposits, including the Paliastomsk Deposit where alluvial-swampy-lagoonal grassy-peat occurs, associated with lacustrine-lagoonal Paliastomi deposits (Fig. 3, drillhole #946-414). In comparison with the marginal alluvial-swampy woody peat type, the distinguishing characteristics of both of these facies types of the cycle in the central peat swamp zone include the simple composition of the peat bed that consist mostly of grassy peats, occasionally with minor amounts of mossy-grassy genetic types and facies. This results from the disappearances of facies OZD and APZ from the facies complex and a sharp reduction in soil formations within these beds. Studies indicated that fine, medium, occasionally coarse sands of facies ARP occur at the base of the various cycle types in the entire area. A single, 0.5-6.5 m thick bed of floodplain deposits occurs in both types upward in the section. It consists essentially of alternating coarse and fine-grained aleurolites, their interlayering and rarely of clays. It is typical of changes between dominantly near-channel (APP) and floodplain interior (APV) facies, with rare participation of floodplain lake and oxbow lake facies. The thickness values of beds of individual floodplain facies in the most complete sections ranges widely (0.25-2.5 m). The most consistent units can be traced for distances of kilometers, even few tens of kilometers. Deposits of individual floodplain- and oxbow lakes may reach thickness values of 1 m. They display lenticular pinchouts within a distance of one kilometer. The upper part of floodplain deposits of both cited categories in the cycle, with rare exceptions, has uniformly been reworked by the swamp vegetation and converted into 0.5 m thick grassy soil and subsoil units (facies PDT). The soil is followed by a peat bed. Unlike in cycles of the marginal peat swamp zone, it is not disrupted by the intrusion of non-organic sediments of various facies, and is composed of low-ash peat of a single genetic category (e.g., bog-grassy, gelinitic-precollinitic) and thus remain uniform (Fig. 2a), or, in other instances, complex (Fig. 2b) [20]. The complex types are composed of various genetic peat categories, each of them represented by an identifiable horizon. Thus drillhole #13 within the 10.5 m thick peat unit in the central peat swamp area of the Imnatsk Deposit displayed these four horizons (upward in the section): (1) peat with woody and grassy-woody gelinitic-precollinitic genetic peat types; (2) sedge-grassy peat with reedy-sedge, gelinitic-telinitic composition; (3) molinitic-sphagnum peat, gelinitic-posttelinitic; (4) Sphagnum imbricatum-peat, at the top of the unit, 3.0-5.5 m, formed above peat swamp level. According to the pollen-spore spectrum of this bed, compiled by Neistadt and Khotinskii [15], the sphagnum peat horizon corresponds to the beech-hornbeam(-linden) phase. According to the Blytt-Sernander climatic classification, this bed correlates with the start of the Subatlantic period. It was dated 2060 ± 150 yr. B.P. (GIN-1993) or 2100 ± 150 yr. B.P. (Mo-251).

The peat beds of the central areas of most deposits are of similar thickness: 5.5-6.5 (7.0), sometimes to 8.5-9.75 (Fig. 4, drillhole #901) and 10.5-12.0 m (Fig. 2b) south of the Nabadsk and in the Imnatsk Deposits, or thin to 0.75-3.2 m in the Lake Paliastomi cycle type (Fig. 3, drillhole #405-415). The start of peat accumulation according to absolute dates occurred at about 5000-6000 yr. B.P., occasionally earlier. This time frame, according to the Blytt-Sernander classification, corresponds to the start of the Atlantic period. The thinning of the Paliastomsk peat bed from 5.9 to 3.2 m was due (Fig. 3) to the reduction in peat accumulation in this part of the future grassy swamp in the interval bracketed by radiocarbon dates 3930 ± 200 (GIN-1234) and 4580 ± 50 yr. B.P. (GIN-1994) and the start of bay development. The bay was connected to the transgressive Black Sea during New Black Sea times. Later, lagoons also formed here. This is confirmed by the almost uniform occurrence of a grayish brown and chocolate brown humus-bearing-sapropelic aleurolite bed over the peat bed. Less frequently, clays with brackish-marine diatoms were deposited in the low-energy parts of the bays (MPZ) that, judging from the diatom biotopes, contained salt waters. Upward in the section essentially sapropelic silt-clay deposit follow, enriched in fresh-water diatoms, but also containing brackish-marine fauna elements that were derived from lagoons (MPL), isolated from the sea. Salinity in the lagoon waters decreased significantly, as indicated by the strong presence of fresh water diatoms. Humus-rich sapropelitic deposits occur upward in the section of the cycle. The grain size remains the same as before but the sediments contain abundant brackish-marine diatoms and marine macro- and microfaunas (forams). This indicates salinity increase in the previously established lagoons. Thus, stages of lesser and high salinity conditions alternated in the Paliastomi lacustrine-lagoonal depositional system. Farthest west (Fig. 3, drillhole #959), the peat bed in the alluvial-swampy-littoral-marine grassy-peat cycle was overlain by fine and medium sands of a higher energy, shallow lagoon (facies MPP). This area was the western continuation of sapropelic silts and clays, deposited in low energy zones of bays and lagoons (MPZ and MPL). Consequently, only in the alluvial-lagoonal

and alluvial-littoral marine, grassy peat (marsh) cycle categories was the peat bed buried under littoral marine deposits in the Paliastomsk Deposit. The littoral marine deposits occur in the most deeply subsided portions of the Potiisk Basin since about four thousand years ago. In the rest of the central peat swamp area, the zone of alluvial-swampy grassy peat cycle development, peat accumulation has continued essentially uninterrupted to the present time. However, research has shown that where the central swamp peat areas are drained by streams, river channel sands overlies peat swamp deposits and the characteristic composition and categories of the cycle have also changed or the cycle type alternates with sediments of another type. Good examples of this are shown in certain central peat swamp locations in the Potiisk, Nabadsk and Pichorsk Deposits. There, the compositional characteristics of sections of the New Black Sea cycle are regularly represented by lateral shifts in the depositional areas of widely distributed sediments; prolonged deposition occurred from the Rioni River and its tributaries. Thus, an alluvial sediment cycle occurred at locations where alluvial-swampy grassy peat (marsh) sedimentations took place (Fig. 1, Drillholes #1M, S-7a, and 221). As the result, during various stages of their existence the peat swamps had a fragmentary character (Fig. 1, drillhole #12 and Fig. 4, drillhole #106).

Cross section X-X' (Fig. 1) and a number of others, not shown here, clearly illustrate the expansion that occurred in the distribution of sections with an alluvial type of cycle and the pronounced westward shift of the peat swamp boundary south of the Nabadsk Deposit, and in conjunction with the shift of its margins in the same direction. As the result, sections in the southern areas of the central woody peat swamp areas of this deposit acquired the character of the alluvial-swampy woody peat cycle category (Fig. 4, drillhole #106 and Fig. 5, drillhole #101) and not of a grassy peat type that is typical of this location and that occurs clearly defined slightly north of Drillhole #901 (Fig. 4) where the influence of the Rioni river was not felt. The change in the type of sediment and peat accumulation cycle is expressed not only by the facies transition from the grassy peat group into a woody category, but also changes in the composition of the bed from a simple to a complex type, that represents a variety of facies. In addition, the bed often becomes subdivided by the appearance in it of quite thick silty-clayey floodplain sediment layers (Fig. 2, drillhole #12). Laterally they occasionally grade into river channel sands. As the cross section indicates (Fig. 4), in drillholes #52r and 5m, the first one-third of the peat bed was destroyed being invaded by the Rioni River channel. Medium and fine-grained channel sands replaced portions of the peat bed. To the east, in drillhole #4m only the lower part of the peat bed was preserved, while in drillhole #17g the peat bed had been completely destroyed and replaced by coarse and medium-grained sand. The medium sand alternates upward in the section with silty-sandy floodplain deposits. Further to the north, on the right bank of the Rioni River this complex displays, along with river channel alluvium in general, a regular facies transition into an alluvial-swampy woody peat cycle. The peat bed has a complex composition here (Fig. 5a, drillhole #101).

With regard to the assumptions of Makedonov and others [10], who assumed that the peat bed is overlain by marine (bar and beach) sands in this part of the Nabadsk Deposit, a correction is required. These sands found in drillhole #5m further to the west are correlatable with a persistent horizon that occurs in drillholes #52r, 28g and 27g. The sands are poorly sorted, contain few heavy minerals, therefore are alluvial in origin according to their genetic characteristics and not bar or beach sands. There are additional examples of changes in the cycle type, including those in the composition of the peat bed in the direction of the river channels. In the central part of the Churiisk swamp on the right bank of the Khobi River (Fig. 5c), the composition of the section in terms of the predominant facies parageneses is very similar to the alluvial-swampy woody-peat cycle type. However, the enclosed peat bed, although of complex composition, is essentially of grassy composition, indicative of deposition near the water's edge of the Khobi River. This is also indicated along a section of the Tsivi River in the Central swamp area of the Nabadsk Deposit (Fig. 5b) where the peat bed is composed entirely of reedy peat, although displaying differences by its complex composition and increased ash content.

Sections of the New Black Sea cycle in the central peat swamp area of the southern margin of the Potiisk Basin are concentrated in the Matakavsk Deposit area. Based on its essential facies paragenesis, the facies strongly differs from earlier described areas, similar in terms of position in the more subsided portion, and is similar in the composition of the alluvial-swampy woody peat cycle sections in the rear margin zone of the peat swamps. Therefore, it also possesses its own characteristic diagnostic features. For example, thick aleurolite beds of old alluvium (Fig. 1, drillholes #943 and 128) occur in the sections in

the floodplain unit under the peat bed. However, the essential difference involves the composition and structure of the peat bed. In addition to the predominance of alder peats in the peat bed, woody-grassy peats of various ash content and precollinitic structure are widespread as are postelinitic and telinitic peat of very high ash content. Grassy peats, so characteristic of the central swamp areas in the most subsided part of the Potiisk Basin, are common, but are relatively rare in sections of the southern border zone. Analysis of the cross sections indicated that the peat units most often is characterized by complex, occasionally very complex composition (Fig. 1, drillhole #108), the result of the alternation of a wide variety of deposits in terms of genetic types and facies, including peat units with a 1.5-8.5 m range of very variable thickness values. There are many such examples. The discussed composition characteristics of this cycle type are essentially related to the position of the central peat swamp area in the raised southern rim area of the Potiisk Basin, directly in contact with spurs of the Guriisk Ridge. They are also related to the development of a dense network of swamp streams. This in general resulted in an efficient drainage of the swamp, often in changes of the peat accumulation regime, the frequent changes in the peat accumulation regime in conjunction with river channel migration and floods. Such conditions were responsible for the conditions necessary for mutual gradation between grassy and woody peats, the occurrence of mineral layers in the old alluvium and occurrence of changes in the peat bed thickness. The earliest peat accumulation here was radiocarbon dated as 5950 ± 100 (GIN-1085).

Based on basic facies parageneses, sections of the New Black Sea sediment and peat accumulation cycle in the central peat swamp zone along the northern margin of the Potiisk Basin (at the base of the Anakliisk Deposit) are essentially similar to sections of the alluvial-swampy woody peaty cycle type at the rear margin of peat swamp zone. However, the enclosed peat bed is of a uniform thickness (5-6.5 m). Its accumulation has started approximately 6000 yr. B.P.

Sections of marine littoral peat deposits of the central peat swamps areas in the Potiisk Basin consist of nearshore marine and continental deposits. These are basically fine, medium, rarely coarse sand, representing beach facies, littoral (MMV) and underwater bars, subtidal ridges, shoals (MMK), shallow, the flat-bottomed surf zone sediments (MMM), underwater deltas and certain other, rare, types, as sediments of distant marine basin areas (MUM) and also medium, rarely coarse-grained sands of river mouths (ARP) and aleurolites in reduced floodplain deposits (APP). Occasionally, humus-sapropelic clays and aleurolites of bay and lagoon facies (MPL) also occur. The predominantly 0.2-1 m thick peat beds, were buried mostly under bar sediments (Fig. 1, drillholes #5ch, 243, 373, 396, 2ch and 223), less frequently beneath river channel sands. They continue on land and grade into various sectors of the peaty section in the central peat swamp zone, discussed before. They pinch out along the seashore. In the present article we are not detailing depositional conditions in the littoral zone. The subject involves a number of pressing problems, associated with Holocene stratigraphy, that require separate treatment, based on our data. A separate paper will be devoted to this topic.

In the Abkhaz-Mergel'sk Basin (northern Rioni Basin), sections in the hilly piedmont and inclined, uplifted plain are lithologically essentially similar to those of the Potiisk Basin where the bouldery-pebbly, pebbly and sandy-silty Holocene cover [2, 9, 19] was formed entirely by alluvial mountain stream, and partly colluvial and talus processes.

Sections of the Colchis Lowland in the Abkhaz-Mergel and Potiisk Basins contain sequence of an incomplete New Black Sea cycle. In these areas the cycle almost always is represented by alluvial facies. Toward the south, it is underlain by medium river channel sands with pebbles, clearly associated with processes of the Inguri. To the north, in the region of mountain valley streams (Okumi, Galizgi, Moskvyy and Kodori Rivers), the channel sediments become coarser as coarse sand, often pebbly-gravelly matter appears in the sections. Alternating layers of coarse, less frequently fine-grained aleurolite, even more rarely sand and clay, deposited in mostly of near-channel floodplain facies occur over the channel alluvium. The deposits are recognizable over great distances. They are 2-4 m thick, rarely thinning to 0.5 m. In the 70 km wide coastal area between the Inguri and Kodori rivers, the alluvial character of this cycle has been locally preserved as it forms the Black Sea shore. More frequently it is replaced by alluvial-swampy woody peat type of the cycle in the area of the Gagidsk and Zоргatsk peat deposits. The base of the peat deposits, 1.0-2.5 m thick, here were dated 2857 ± 50 yr. B.P. (GIN-647) and 4150 ± 40 yr. B.P. [16]. In other parts of the

Black Sea coast, the peat beds are absent from the littoral sequence. In addition to aleurolitic floodplain deposits, the facies complex includes sediments of facies APZ, sapropelic fresh water lake deposits (OSP), in combination with soil formations (PDD) and facies OZD, as clays with alder-bearing gelinitic-posttelintic peat. The lower portion of the cycle is composed of fine- and medium sand. The facies group in general represents the alluvial swamp interval of the cycle, displayed in the coastal exposures at the town of Ochamchir. Further westward, it was partly destroyed and appears also on the Black Sea floor in the nearshore zone. The deposits were tracked on the seam bottom as far as they were accessible, for a distance of ≈ 100 -150 m. Transgression covering land areas in the Abkhaz-Mergel' Basin and shore erosion took place along the total length of the coastline between Inguri and Kodori. Matkava [12] noted coastal recession rates of 4-5 m/yr over a period of forty years. Comparisons of the changed composition and structure of sequences in various types of Colchis Lowland sections from east to west in the Potiisk and Abkhaz-Mergel'sk Basins indicate that sections in the Abkhaz-Mergel'sk Basin correspond to three morphological zones of sedimentation. These were the rear margin of the Colchis Lowland, the marginal zone of peat formation (alluvial, alluvial-swampy cycles) and the peat swamp margin (alluvial-swampy, woody peat cycle). Clearly, peat deposits in the central peat swamp zone (alluvial swamp, grassy peat cycle), so widespread in the Potiisk Basin, are completely absent from the Abkhaz-Mergel'sk Basin on the continent. Probably they were partly eroded and buried under Black Sea bottom deposits at significant offshore distances, as shown by the Ochamchir example.

The Kobuletsk Basin includes the southern part of the Colchis Lowland. Its width and the surrounding zone was significantly narrowed by the proximity of the Black Sea shore and the Adzhar-Trialetsk Mts. that locally extend to the shoreline. As the result, the E-W width of the Lowland is only 10 km along the Natanebi and Choloki rivers [17]. It is associated with the peat area with the Kobuletsk peat complex of Ispani in the Choloki-Kintrishi interfluvial area that stretches along the seashore, separated from it only by a littoral ridge, 4.5-10 m high.

As the area surrounding the Lowland was not included in the present research, its detailed lithological and facies characteristics are not discussed in the present paper.

The thickness of the Holocene deposits of the Colchis Lowland vary between 21-46 m. Our study of strike and dip sections and utilization of Tvalchrelidze's conclusions [17] on this area indicate that the base of the Holocene that corresponds to the Old Black Sea stage almost everywhere is part of various sections of very poorly sorted clayey aleurolites, with pebbles, coarse and fine sands. Pebble sizes range between 2-4 cm. Sandy-clayey and sandy cement occurs rarely between the gravels. Such a wide grain size range in the sediments and the total absence of sorting indicates that the sediments were deposited by mudflows with rock fragments, derived from thawing glaciers [1]. These Old Black Sea stage deposits are 4-17 m thick and thicken toward the basin center. They overlie bouldery-gravelly mountain stream channel deposits and their alluvial fans (facies ARG) along the Black Sea. The Karangatsk marine units with typical marine fauna occurs at stratigraphically lower horizons. Thus, even if during Old Black Sea times large areas of the Potiisk Basin (with a center 13-15 km inland from the seashore) were covered by marine sands, then according to our and Tvalchrelidze's earlier [17] data, it is also occupied by widespread, exclusively continental deposits. Upward in the section, deposits of the Old Black Sea stage, as in the Potiisk Basin, alternate with sediments and peats of the single, incomplete New Black Sea cycle. Its thickness ranges between 25-34 m. Deposition is estimated to have started 7000 yr. B.P. or slightly earlier. Here and in the rear zone of the Potiisk Basin, as well as in the corresponding zone of the Kobuletsk Basin, sections of the New Black Sea cycle are represented by the alluvial facies type of the cycle. However, the alluvium is not of the alluvial plain type but mountain alluvium. The base of the cycle consists of bouldery-gravelly mountain river stream channel sediments (facies ARG), locally of piedmont alluvial fans (PPK), of unknown thicknesses. Grain size decreases upward in the section. East of this zone, with the exception of the southernmost area, the upper part of the section is almost entirely represented by coarse and fine sand, with pebbles and gravel, occasionally coarse aleurolites, with channel pebbles from mountain streams (ARP). Channel sands and coarse aleurolites display facies displacement by floodplain clays and fine aleurolites (most often, of facies APV) westward, in the central part of this zone, as far as the peat accumulation area, and as far to the north as the littoral margin. In the southern margin area the bouldery alluvial fan deposits of the piedmont parts of the section westward grade into clay, with pebbles

from proluvial wash. Farthest south, the southern rim of basin, the entire New Black Sea cycle through the most recent times consist of bouldery-gravelly deposits. These represent the Kintrishi River mountain alluvium.

Farther westward, in the area of peat accumulation, in the area of the deepest subsidence of the Kobuletsk Basin, the New Black Sea sediment-peat accumulation cycle is represented essentially by two facies cycle types: an alluvial-swampy and a mostly alluvial-lacustrine-swampy facies, with moss peats (bog peats). Both differ from the alluvial type by a complex facies paragenesis of the bed, deposited above mountain river channel alluvium, that is located at the base of the cycle. At the same time it displays a facies relationship with the alluvial type both east- and northward. The composition of the alluvial-swampy cycle category is essentially similar to that located in the corresponding zone of the Potiisk Basin. However, it differs in its significant granulometry of the enclosed sediments that formed in mountain stream alluvium. This cycle type includes sediment sequences at the base of the northern part of the peat accumulation area. From the center southward, in the area of peat swamps it is represented by an alluvial-lake-swamp moss-peat (bog) cycle type. It contains bouldery-gravelly, occasionally pebbly mountain river channel alluvium at its base, manifested in both facies types. Upward in the section, the bouldery-gravelly deposits usually grade into coarse-grained sands of coastal plain river channels. Farther downstream, fine and coarse floodplain aleurolites occur or only coarse aleurolites, without sands. This complex is $\approx$ 5-6 m thick in the cycle.

Sapropelic lake aleurolites, overlying the floodplain aleurolites, locally clays with fresh water diatoms (facies OSP) always occur in this cycle higher in the sequence. This unit is 1-6 m thick. Only in the western sections do brackish, occasionally brackish-marine diatoms appear (facies OSS, OSO; Fig. 1, drillholes #381, 379 and 384). The upper portion of these sediments have been reworked over large areas into woody soil matter (facies PSD). The cycle over their entire area is topped by a usually 6-8 m thick peat bed that thins to 4 m toward the margin of the peat swamp area. It is of complex composition; it consists of various genetic peak categories, representing various facies. Three horizons were present; a lower, middle and upper. In addition to specific material components, we have utilized research data by Dokturovskii [5] to characterize them. The 0.5-4 m thick lower horizon consists of gelinitic-precollinitic alder peat that contains ash in the peat swamp margins. Occasionally it was mildly decomposed (posttelinitic peat structure) but in the central areas it is often replaced by clay with peat of the muddy peat swamp facies (OZD) in which aleurolites and clays of overgrown water bodies of forested floodplain (facies APZ, drillhole #387), occasionally soil horizons (facies PDD, drillholes #379, #381), also participated. Thus, the lower part of the bed is always of complex composition. The middle part of the bed, 0.5-2 m thick, consists of gelinitic-telinitic reed, mostly of rootstock and sedge. Various amounts of alder also occur, occasionally even sphagnum. The peat is always of low ash content, without terigenous sediment layers and thus the horizon is of a uniform, simple composition. The upper, 3-5.5 m thick horizon is represented by gelinitic-telinitic, sphagnum-bearing peat with low ash content, forming "cupolas", elevated over the swamp surface, excepting certain marginal zones, almost in the entire area. The source matter of peat contains predominant amounts of Sphagnum imbricatum. Its widespread distribution, according to the pollen-spore diagrams by Dokturovskii is attributed here and in the Imnatsk swamp of the Potiisk Basin to the Subatlantic period of the Blytt-Sernander scheme. These diagrams show that peat accumulation in the Kobuletsk Basin started later than in the Potiisk Basin! approximately during the second half of the Atlantic period. Absolute ages date peat at the base of the bed (drillhole #382) 4480 ± 200 yr. B.P. (GIN-1133) and (drillhole #387) 5000 ± 300 yr. B.P. (GIN-862g).

The following conclusions are warranted:

1. Study of the Holocene sections of the Colchis Lowland by detailed lithofacies analysis, based on a large number of recently acquired drill samples revealed new aspects of the composition of the deposits. These study methods were applied for the first time in this subject area. The sediment composition characteristics clearly defined the incomplete New Black Sea cycle of deposition that started $\approx$ 7500-7000 yr. B.P. Various facies parageneses established facies cycle type in the sections and their distribution pattern laterally. The Colchis Lowland deposits have been subdivided into the following zones: the rear margin of the Lowland, the margin of the peat formation zone, the peat swamp margin area, the central peat swamp areas, their basin-margin zones, and the littoral margin of the peat accumulation area. Thick beds (maximum 10 m) of river channel with floodplain alluvium over it occurs in

the section in the lower half of the cycle of facies types. These beds in these sediment zones continuously overlie an erosional surface, cut into lower Old Black Sea sand and aleurolite sequence in the past accumulation area. Over this surface the peat swamps develop with their thick (maximum 12 m) peat bed. This fact confirms that peat formation in the Lowland preceded the continental deposits, contrary to the general opinion of other workers [4, 10, 14, 15, 19]. These considered them lagoonal-marine, lagoonal, lacustrine, occasionally beach deposits. Recently, Dzhanelidze expressed the view in his monograph that "the study of the stratigraphy of littoral peat bogs and the underlying marine deposits in our view indicate that the peat swamps essentially formed at locations, previously occupied by Old Black Sea lagoons, as the result of gradual expansion of the lagoons" [4, p. 46]. Tsereteli has shown that "the peat (in the area of Lake Paliastomi) is underlain by grayish-green clays and muddy sands of lacustrine-lagoonal origins" [19, p. 401]. The sands that underlie the peat bog in the Nabatsk Deposit have been considered of delta-subtidal bar, occasional beach origin by Makedonov and Ishina [10]. Based on spore and pollen diagrams, Neishtadt and Khotinskii [15] considered the muds that underlie the peat beds (in the Imnatsk Deposit - L. B.) products of a lagoonal stage.

2. Based on the lithofacies composition of continental deposits that directly underlie the peat bed, it is clear that the deposits are basically of floodplain origin in the Potiisk Basin. In the Kobuletsk Basin they are mostly clayey deposits, with an abundance of fresh water diatoms that lived in sparopelic lakes. The sediments here occur among lowland floodplain aleurolites that underlie the peat bed only at the southernmost location. Consequently, two basic sediment paragenetic categories have been identified in the peat accumulation area of the Colchis Lowland directly beneath the peat bed: a "floodplain" complex, typical of the northern half of the Colchis Lowland, and a "lacustrine" type, that occurs in the southern half. These facts refute Tvalcherlidze's assumption [17] that the peat bed in the Kobuletsk Basin is underlain by nearshore shallow marine deposits. The geological setting of units that directly underlie the peat in the Abkhaz-Mergel'sk Basin has not yet been sufficiently documented; the basal portion of the peat accumulation area there is buried by transgressive Black Sea sediments, as shown earlier in the present paper.

3. Studies have shown that the westward shift of the sedimentation zone from the rear margin of the Colchis Lowland was accompanied by a clearcut change in facies parageneses in the upper half of the New Black Sea depositional cycle. This involved a shift from typical alluvial, toward peat swamp environments. Change in the peat bed involved a composition change from complex to simple, a steady increase in the bed's thickness, toward the peat swamp center, a simultaneous replacement of the woody genetic peat group by grassy (marsh) peat, with mossy "cupolas", in the most subsided portions of the Potiisk and Kobuletsk Basins.

4. However, these naturally evolved natural relationships had been disrupted in areas where the region's sedimentation zones were subdivided by floodplains of streams that traverse the region on their way to the Black Sea. The disruption manifested itself in the partial or total erosion of the peat bed, in a change of the compositional character, previously well defined for a given type zone, a change in thickness of the enclosed genetic peat group, and lateral changes of the sedimentation zones themselves within the area of peat accumulation. These zones occasionally displayed interruptions in given facies, associated with formation of the peat bed. Morphologically, the large transverse streams subdivided the peat unit into individual lenses. Each lens occupied a given interfluvial area in which peat accumulated as far downstream as the Black Sea littoral.

5. The identification of the patterns of sediment development on the Colchis Lowland shed new light on the evolution of sediment accumulation during Holocene, the last stage of Earth history in the subject area, and simultaneously characterize the model of recent, deltaic type peat accumulation in the intermontane setting of the Rioni Basin that is open toward the sea. These principles also provide means with the help of which peat may be efficiently exploited and an assistance in establishing its commercial potential.

LITERATURE CITED

1. L. I. Botvinkina, "Layering in sedimentary rocks," Trudy GIN, Academy of Sciences of the USSR, No. 59 (1962).
2. E. K. Vakhaniya, "Geology of the Colchis Lowland and its hydrocarbon content," Trudy VNIGNI, Gruzian Division, No. 151 (1973).
3. P. D. Gamkrelidze, "Tectonics of the Colchis Lowland," Geol. SSSR, 8, Nedra, Moscow, 327-338 (1964).

4. Ch. P. Dzhanelidze, The Paleogeography of Gruzija (Georgia) during the Holocene [in Russian], Metsniereba (1980).
5. V. S. Dokturovskii, "Data on Transcaucasian peat studies," Pochvovedenia (Soil Science), No. 2, 180-201 (1936).
6. N. V. Dumitrashko, "Caucasus," in: The Mountain Region of the European USSR, and the Caucasus [in Russian], Nauka, Moscow, pp. 90-226 (1974).
7. Yu. A. Zhemchuzhnikov, "Coal-bearing beds and their study methods," Zap. LGI, No. 25, 2, 23-46 (1951).
8. Yu. A. Zhemchuzhnikov, V. S. Yablokov, L. I. Bogolyubova, and others, "Composition and deposition conditions of sediments of Donets Basin Middle Carboniferous coal-bearing formations and coal beds," Part 2, 15, Izd-vo Akad. Nauk SSSR, 340 (1960).
9. A. G. Laliev, "Problems of the geotectonic nature and geological evolution history of the Colchis Lowland," Trudy GIN, Akad. Nauk SSSR, 10 (15) (1957).
10. A. V. Makedonov and T. A. Ishina, "The Colchis Lowland," in: Principles of the Composition and Formation of Coal-Bearing Units and Methods of Forecasting Coal Content [in Russian], Nedra, Leningrad (1986).
11. D. I. Mamaladze, The Marine Pleistocene in the Colchis Lowland [in Russian], Metsniereba, Tbilisi (1975).
12. D. I. Matkava, "Shore erosion in the northern Colchis region," Soobshch. Akad. Nauk SSSR, 2, No. 1, 5-9 (1976).
13. E. E. Milanovskii, Geology of the Caucasus [in Russian], Izd-vo Akad. Nauk SSSR (1963).
14. A. V. Motsereliya, Transformation of the Colchides [in Russian], Izd-vo Akad. Nauk SSSR (1954).
15. M. I. Neishtadt, N. A. Khotinskii, A. L. Devits, and N. G. Markova, "Imnatsk Swamp (Gruzian SSR)," in: Paleogeography and Chronology of the Upper Pleistocene and Holocene, Based on Radiocarbon Dating [in Russian], Nauka, Moscow (1965), pp. 105-112.
16. V. P. Sluka, "Palinological and lithofacies studies of Holocene peat deposits of the Colchis Lowland," in: Palinography of the Holocene and Marine Palinology [in Russian], Nauka, Moscow, p. 53-57 (1973).
17. M. G. Tvalchrelidze, Lithology of Recent Sediments in the SE Black Sea Coastal Region [in Russian], Candidate Dissert., Geol.-Mineral. Sciences, GIN Akad. Nauk Gruzian SSR, Tbilisi (1987).
18. P. P. Timofeev, "Geology and facies of the Jurassic Formation of Southern Siberia," Trudy GIN Akad. Nauk SSSR, No. 197 (1969).
19. D. V. Tsereteli, Pleistocene Deposits of Gruzija (Georgia) [in Russian], Metsniereba (1966).
20. V. S. Yablokov, L. I. Bogolyubova, and L. P. Nefed'eva, "Composition of coal beds and coal types of the Erunakovsk Formation, Kuznetsk Basin (Kuzbass)," Trudy GIN Akad. Nauk SSSR, 136, No. 3 (1951).

THE ROLE OF THE SUBJACENT COMPLEX IN BAUXITE-FORMING SOUTH URAL DEPOSITS

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UDC 553.492(450.5)

Interrelationships of the rocks of the bauxite-bearing Orlov suite of the Upper Frasnian with the subjacent rock complex in South Ural are studied. The formation and evolution conditions of this complex are described. Attention is focused on lithologic and facies features of limestones of the Samsonov suite of the Upper Frasnian, occurring at the base of bauxite deposits in South Ural. The facies zones of the bauxite-bearing Orlov suite have been found to be associated spatially with specific facies zones of underlying limestones. The prebauxite topography was of a karst-erosional nature; a necessary condition for development of this topography is low-angle ancient elevations.

Late Frasnian (Orlov) continental discontinuities are identified in the western slope of the South Ural. It is associated with bauxite formation. There are also regions with continuous Upper Frasnian sediment sections (Fig. 1). The Late Frasnian continental interruption was developed within arch rises of the Chusovskaya zone of the western slope of the Urals (shallow-sea shelf area). On the other hand, the Sergin subzone (the barrier reef region), the Mikhailov subzone (the deep shelf) of the Bel'sk-Elets'k zone, and a considerable part of the Chusovskaya zone were areas of uninterrupted sediment accumulation [4].

An extensive ancient land of the Orlov time which included South Ural bauxite deposits (SUBD) (stretching some 150 km latitudinally) was situated in the northern part of South Ural within peripheral uplands of the Bashkirian arch system. The land area associated with the Nugush upland occupies a much smaller territory (approximately 40 km across).

An influence of the basement topography and the composition of the subjacent rock complex on bauxite quality and thickness and facies of bauxite-bearing horizons has been mentioned by previous investigators for Paleogene bauxites of Kazakhstan [7] and Devonian bauxites of North Ural [5].

The present paper is an attempt at establishing similar regularities for bauxites of the South Ural bauxite mines (SUBD).

Lithologic and Facies Zonality and Formation Conditions of the Subjacent Rock Complex

Within the Bashkirian paleoland, Precambrian, Upper Givetian, and Lower Frasnian rocks occur at the base of the bauxite-bearing Orlov suite; there are also Upper Frasnian sediments of the Samsonov and Mendym suites. The distribution and succession of different-aged rocks at the base of the Orlov suite are largely controlled by local paleostructures.

The Precambrian underlies the Orlov suite in the central parts of the Bashkirian upland. It is undercut by rare bore holes east of Karatu and north of the Suleimanov brachyanticline (see Fig. 1). Most rocks are schists, silt, and sandstones of the Vendian Ashinsk suite; ancient dolomites, limestones, siltstones, and sandstones of the Karatu series of the Upper Riphean are less frequent.

The Upper Givetian and Lower Frasnian sediments (Cheslav and Sergaev suites) underlie the Orlov suite in the eastern margin of the Chusovskaya zone in the southwestern and northern wings of the Porozen brachyanticline.

The Upper Givetian (Cheslav) sediments are represented by gray layered organogenic limestone; the Lower Frankian (Sargaev) rocks are gray clayey striated limestone and dolomite.

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The Lower Frasnian sediments (the Domanik suite) surround in a broad band on the west, north, and east the outcrops of the Cheslav and Sargaev suites, represented by dark-gray clayey bituminous limestones and clayey schists.

The Upper Frasnian Sediments of the Samsonov and Mendym Suites. The Samsonov suite is widespread in SUBD and its flanks: the eastern wing of the Suleimanov and southwestern wing of the Porozhen brachyanticlines, in the Uluir syncline, and near Karatau. The Samsonov suite consists of dense light fine crystalline and recrystallized organogenic clastic limestones. These limestones are widespread in the middle part of the suite, in the territory's western area, near Karatau.

The Mendym suite is common west of the Suleimanov brachyanticline. This is the only location in South Ural where the roof of the Mendym suite evolved during the Orlov continental interruption. The Mendym suite there consists of gray, less often light- and dark-gray organogenic-clastic limestones formed in a semiinsulated sea basin (lagoons). To the west and east, Mendym limestones give way to facies pattern of some Samsonov rocks.

The Samsonov suite near SUBD is represented by dense light-gray, gray, or pink-gray massive fine crystalline and sphanitic organogenic-clastic limestones. They cover a small territory (25 × 10 km) and are macroscopically homogeneous, because, mostly, they are recrystallized. Attempts at separating these rocks into the common classifications of lithologic types had failed. We resorted to the method proposed by Wilson [13]. Modern, so-called energy classifications of limestones by U.S. geologists, recommended also in [10], are mostly based on the quantity of original silty carbonate matter (micrite) and grain types. It is believed that for describing the water circulation conditions it is more important to estimate the concentration of calcareous oozy filler matter than to determine the grain sizes and shapes. Wilson distinguished 20 microfacies and defined the types of facies settings from barrier reef to lagoonal and littoral environments according to the predominant presence of some of these microfacies.

Lithologic varieties (microfacies) of limestones studied here are described in the sequence of representativity and area of occurrence. The microfacies were used as a criterion to determine facies belts.

1. Biomicrites, bioclastic wackestones,* or microcrystalline organogenic clastic limestones are most widespread and account for some 60% of the limestone types studied. Mostly, they are confined to the southeastern wing of the Uluir syncline.

Faunistic fragments mixed with burrowing organisms are widespread; they are contained in a micritic (<0.005 mm) filler. The fragments are poorly sorted or unsorted. According to the relative concentrations of fragments and cement, unsaturated (<50% of fragments) and saturated (>50% of fragments) biomicrites are distinguished. Coarse- (1-3 mm) and small-detrital (0.1-1 mm) varieties are separated among fragment size groups. Fine detritus (<0.1 mm) is found in both varieties. Organic detritus is represented by fragments and whole shells of brachiopods (agripids), crinoids, gastropods, tentaculites, ostracods, algae, bryozoa, and foraminifers. In the southeastern part of the area, saturated biomicrite contains semi-rounded quartz grains (up to 5%), up to 0.1 mm in diameter. Such limestones are deposited in shallow-water neritic conditions with active and weak water movements and free water exchange.

2. Biomicrosparites and biosparites are lithobioclastic, peloid, or organogenic-clastic limestones with sparitic (0.2 mm) or microsparitic (0.005-0.02 mm) cement (25% of all limestone types). They are mostly found in the Kuksha bauxite-bearing strip. Microsparitic varieties with isolated organic remains and no lithoclasts or pelites are also isolated; there are also varieties variously saturated with organic remains.

Lithobioclastic and peloid sparites and microsparites (intrasparites) are common. The bulk consists of small (0.2-0.3 mm) poorly rounded and angular lithoclasts and fragments of crinoids, ostracods, seed chambers of algae, and pellets enclosed in sparitic or microsparitic fine crystalline (up to 0.05 mm) calcitic mass. Virtually every fragment is surrounded by a 0.1-mm fringe filled with calcite crystals; some cement crystals exhibit a centripetal orientation. The composition of organic detritus is similar to that of biomicrites. According to Wilson [13], rocks accumulated in an intense hydrodynamic environment under constant wave action near their active basis; as a result, the pelitic ooze mass was washed out. Poorly

*Here and hereafter, we use Wilson's terminology [13].

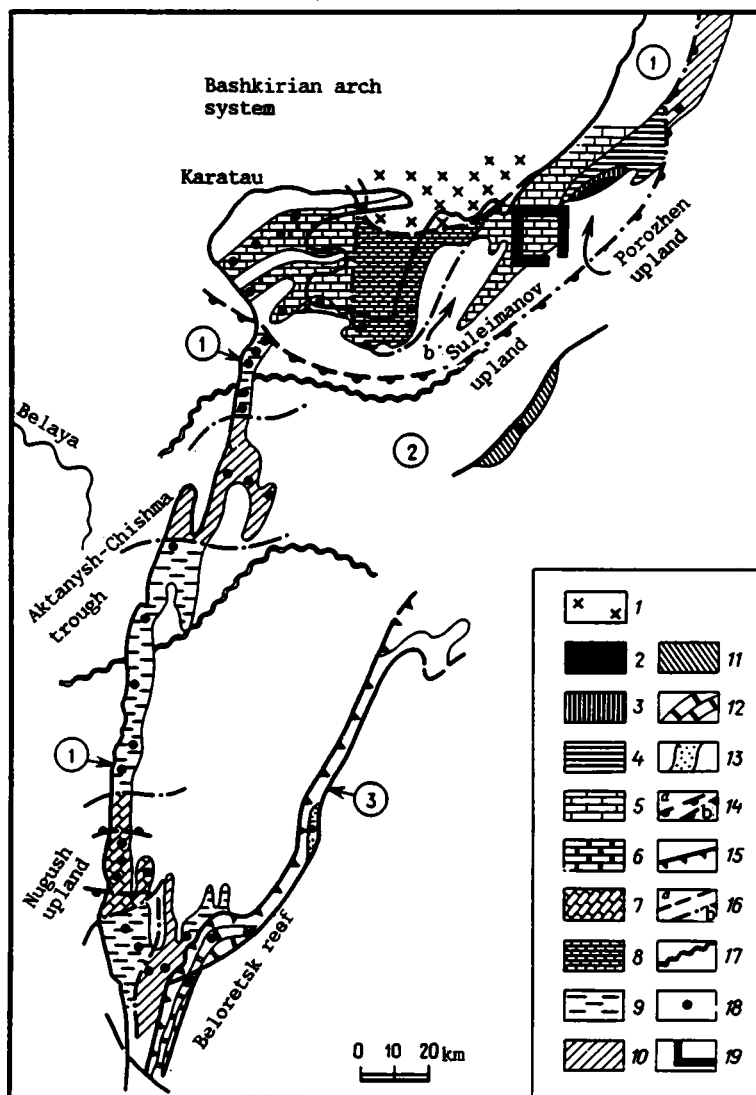


Fig. 1. Lithologic-facies map of the base of the Orlov bauxite-bearing suite (modern data): 1) Precambrian sediments; 2-4) Upper Givetian and Lower Frasnian sediments (suites: 2 - Cheslav, 3 - Sargaev, 4 - Domanik); 5-9) Chusovskaya zone (5 - Samsonov suite limestones, 6 - same, reefogenic, 7 - silicic carbonate sediments (lower section) and light stratified limestones (upper section), Samsonov suite, 8 - gray layered limestones of the Mendym suite, 9 - gray layered partly clayey and bituminous limestones of the Mendym suite); 10-11) Mikhailov subzone, Belsk-Eletsk zone (10 - dark-gray bituminous limestone, silicic-clayey schists of Mendym and Manticoceras suites, 11 - silicic-clayey schists, argillites, and dark-gray clayey limestones of the Mendym suite); 12-13) Sergin subzone, Belsk-Eletsk zone (12 - gray layered limestones, 13 - finely layered light limestones of a reduced thickness); 14-17) boundaries: 14) occurrence area of Orlov continental discontinuity (a) on the segment where it coincides with the boundaries of the Chusovskaya and Belsk-Eletsk zones (b), 15) Sergin subzone (on overthrust), 16) between different-aged sediments (a), lithologic-facies boundaries within *Manticoceras intumescens* Beer zone (b), 17) Aktanysh-Chishma trough, Russian Platform; 18) reference sections (except for SUBD); 19) SUBD area (see Fig. 2). Circled numbers: 1) Chusovskaya zone; 2-3) Belsk-Eletsk zone (2 - Mikhailov subzone, 3 - Sergin subzone).

rinsed lithobioclastic pelosparites are also found sometimes; the cement of these units partially consists of micritic mass. These rocks are a transitional variety between the above two microfacies.

3. Algal stromatolitic microgranular limestones. These rocks are infrequent (5-10%) and mainly developed near the axis of the Uluir syncline and its northwestern wing. The

thin layers grow together densely; the finer-grained calcite is more often formed on higher-lying areas, producing a stratification that does not conform to a gravitational pattern. According to Wilson [13], such a stromatolitic structure is typical for the tidal zone.

Based on the predominant area development of the preceding microfacies of Samsonov limestones on the territory SUBD and its flanks, a facies zonality (facies strips) has been identified. Three zones (types) of marine facies settings are distinguished (from west to east, from the shoreline of the Samsonov time; see Fig. 2): 1) shallow littoral zone with traces of tidal activity; 2) shallow neritic zone with active and weak movements and free water exchange; 3) shallow zone with intense hydrodynamics (a setting of wave zone shoal).

These zones stretch in a northeasterly direction. Their boundaries are parallel to the boundary of shallow and deep shelves of the Chusovskaya zone and the Mikhailov subzone of the Bel'sk-Elets'k zone. The above data confirm a regional environment established in [3] for the beginning of the Late Frasnian in this territory and differentiates facies zonality in greater detail. The carbonate body of Samsonov limestones within these limits and the above facies zonality should have a thickness consistent laterally and in the seaward direction (from west to east). Presumably, a slight increase of thickness exists in the intense hydrodynamic zone: in this environment carbonate structures could be formed in the direct vicinity of the area where shallow shelf gives way to deep shelf.

Mendym limestones with a washed-out roof were studied on 40 transparent polished sections collected near the western wing of the Suleimanov brachyanticline. Several limestone varieties were determined which had no regular patterns in area distribution.

4. Lithobioclastic coarse detrital peloidal biomicrite-biomicrosparticle (or behind-reef calcispheric wackestone). Spherical formations (calcispheres) which probably are seed chambers of algae are present in many of the Mendym limestones, indicating a lagoon setting.

There are varieties with organogenic detritus of a limited set of species, such as ostracod microsparites and gastropod micrites. The absence of any other organic remains in these specimens is evidence that sediments were deposited in conditions of a behind-reef lagoon space.

Peloids are found in virtually all rock varieties. There are also pelosparites made up of pellet accumulations. This indicates active reworking of calcareous sediment by burrowing organisms, which can only occur in calm shallow basins.

The above varieties of Mendym limestone were accumulated in a shallow setting of a semi-insulated lagoon at the back of the Samsonov carbonate structures. The occurrence area of Mendym limestone can thus be interpreted as a lagoon at the beginning of the Late Frasnian. The Mendym suite contains algal stromatolitic limestones with an antigravitational stratification similar to that described for the Samsonov suite. This suggests similar formation conditions of both types of rocks in a littoral-marine tidal zone.

The limestones of the Mendym suite, which are widespread near the Mikhailovsk subzone, where no traces of the Orlov continental interruption are observable; are a relatively deep-sea marine deposit, largely similar to limestones of the underlying Domanik. Mostly, these are clayey and micritic rocks with organic remains of a wide set of species. The Mendym suite also contains argillites and clay-silicic schists. Thus, Mendym suite sediments even in the Mikhailovsk subzone accumulated in settings substantially different from those in the western wing of the Suleimanov brachyanticline.

Formation conditions of Mendym-Samsonov limestones on the remaining territory of South Ural are discussed below.

Near the Nugush rise the Mendym horizon has a bipartite structure: the lower part is represented by depressional dark clay-carbonate sediments, similar to Domanik; the upper part is represented by light-gray organogenic-clastic limestone.

In areas where the Upper Frasnian is continuous, a certain lithologic-facies zonality of the Mendym-Samsonov sediments can be observed [6, 12].

In the western wing of the Bashkirian anticlinorium in the Chusovskaya zone between the Bashkirian and Nugush paleolands, the Upper Frasnian level includes, virtually everywhere, gray and dark-gray stratified partly bituminous limestones of the Mendym suite.

At the beginning of the Late Frasnian, this area was the lowest-elevation part of the shallow-sea Chusovskaya shelf.

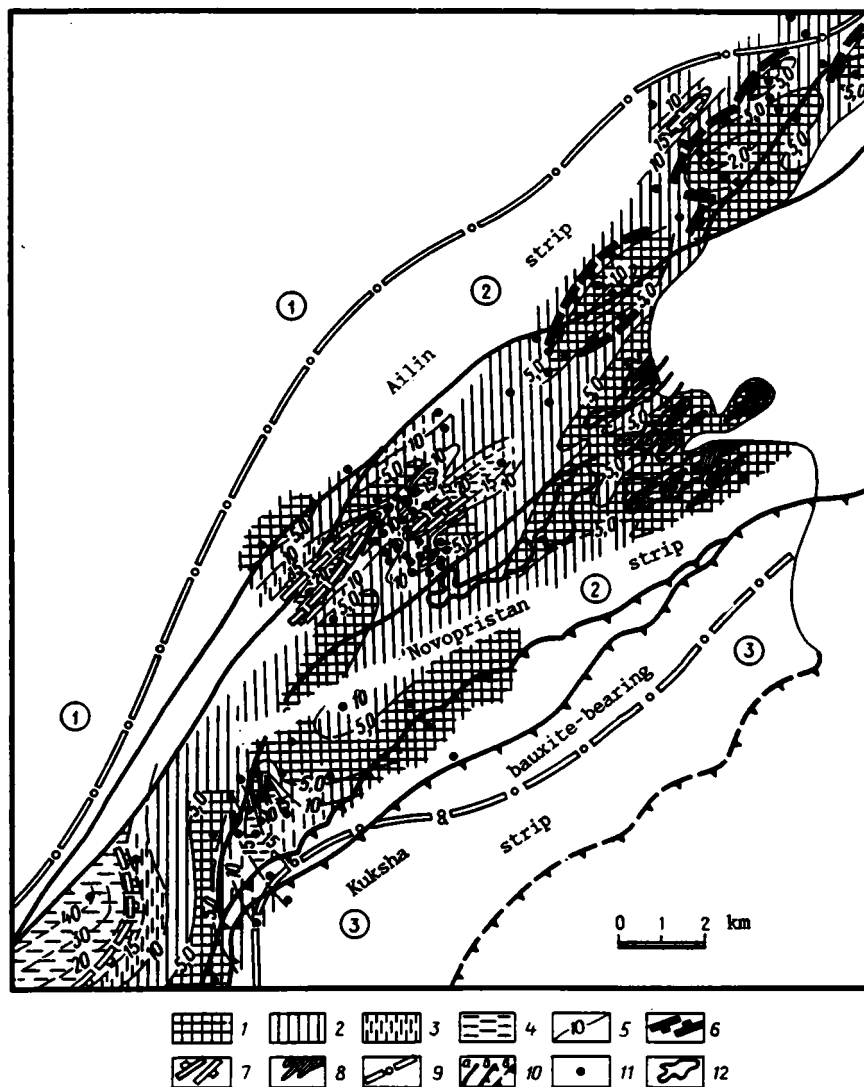


Fig. 2. Map of isopachytes of Samsonov suite with facies zonation elements (on palinspastic base): 1-4) Samsonov suite limestone thickness, m: 1) 5, 2) 5-10, 3) 10-15, 4) >15; 5) isopachytes in meters; 6) boundaries of bauxite accumulation favoring areas with limestone thickness <4-5 m; 7) boundaries of areas unfavorable for bauxite accumulation with limestone thickness >18-20 m; 8) outcrops of Domanik sediments at the base of the Orlov suite; 9) boundaries of facies zones; 10) main faults (a - downthrows and displacements, b - overthrusts, c - overthrust boundary of Kuksha allochthonous plates in palinspastic reconstruction); 11) benchmarks of isopachytes (wells that penetrated the full thickness of the Samsonov suite); 12) areas of thick red bauxites. Circled numbers (zones, types) of facies settings of limestone accumulation (by microfacies): 1) shallow-sea littoral, 2) shallow-water neritic zone with active and weak water motion and free water exchange, 3) shallow zone with intense hydrodynamics: wave zone shoals.

At the center of the strip of Mendym sediments, identified in the modern structure in the western wing of the Bashkirian anticlinorium, a contemporaneous entity of the Mendym suite - the Mantikotserov suite - is common at the latitude of the Aktanysh-Chishmin trough of the Hama-Kinel system of the Russian platform [9]. The Mantikotserov suite consists of deep-sea dark-gray and gray bituminous limestones and marls with silicic-clay schist interlayers. These deep-sea sediments on a sublatitudinal structure transverse to the Ural zone indicate that during the Late Frasnian the deep-sea troughs of the Kama-Kinel system converged with the deep-shelf of Ural (Mikhailovsk subzone of the Bel'sk-Eletsk zone).

The Mendym suite, which is common in the Mikhailovsk subzone, consists of deep-sea dark-gray clay limestones interstratified with dark-gray and black argillites, siltstones, and clayey-silicic (sometimes radiolarian) schists that have been preserved in the modern structure northeast of the Porozhen brachyanticline and in the Yuryuzan syncline.

The Sergin subzone in south Ural near the Beloretsk-Tirlyan segment, where no Late Frasnian Orlov interruption took place, has age analogues of Samsonov and Mendym suite light-gray and gray layered limestones linked by gradual transitions with underlying and overlying rocks. East of these limestone areas there are light nearly white rhythmically interstratified thin-layered limestones formed in the eastern slope of the Beloretsk reef that faces the continental slope.

The carbonate body of Samsonov type with the roof washed out during the Orlov interruption, which is well developed in the southeastern periphery of the Bashkirian arch, is the only bauxite-bearing object in South Ural. It is located in the eastern part of a shallow shelf. On one side it is bounded by deep (Mendym and Domanik) sediments. On the other it is bordered by shallow lagoon (Mendym) sediments and consists of massive organogenic-clastic limestones with specific facies zonality.

In order to determine the correlation of lithologic facies zonality of the Mendym-Samsonov and underlying sediments with the formation history of Samsonov-type carbonate bodies, data on lithologic-facies composition of Domanik deposits were analyzed. Materials of Domrachev [6] and Tyazheva [12] were used extensively, as were specimens collected by the authors in the SUBD area and its flanks and the Beloretsk reef.

Most investigators [2, 8, 11] interpret the Domanik suite as marine deep-sea sediments corresponding to the maximum transgression at the end of the Early Frasnian. During the Domanik time, the entire territory of the miogeosyncline in the South Ural area experienced considerable subduction.

In the most shallow areas compared with other Domanik facies, the limestone facies was deposited. It has been documented in eastern parts of Karatau and the wings of the Suleimanov and Porozhen brachyanticlines; it is represented by dark-gray clayey and bituminous limestones interstratified with calcareous argillites and schists.

The eastern part of the Bashkirian system of arches, including the Suleimanov and Porozhen paleouplands that existed since the Orlov time, was a positive structural element as early as in the Domanik time.

The Nugush paleoupland, formed in the accumulation area of deeper normal Domanik facies, began to operate as a positive paleostructure as late as during the second half of the Mendym-Samsonov time.

This can be the factor responsible for uneven rise, separation, and erosion of the land topography formed in place of these structures during the Orlov continental interruption. The inheritance of Mendym-Samsonov facies can thus be viewed as a factor of an Orlov bauxite accumulation. It is definitive for the development of carbonate bodies of Samsonov type on earlier-founded positive structures.

Role of Underlying Rock Complex in Distribution of the Facies of the Orlov Suite

The underlying rock complex zones classified by age, structure, and lithologic-facies characteristics are overlain by the multifacies Orlov suite in the Askynov horizon. In this section we describe the correlation of lithologic-facies zones of the underlying complex and the facies subzones of the Orlov suite described earlier [1]. We try to determine the regularities and causes of these correlations.

The Precambrian consists mainly of terrigenous Vendian and Riphean rocks. The strata are overlain by terrigenous rocks of the Orlov suite: sandstone and micrite, sometimes ferruginated.

The Upper Givetian and Lower Frasnian sediments (Cheslav and Sargaev suites) of the northwestern wing of the Porozhen brachyanticline consist of organogenic and clayey limestones. They are overlain by lagoonal dark-gray argillites, marlstone of the Orlov suite that includes at the bottom rare thin lenses of variegated allites of the Porozhen facies subzone.

The Lower Frasnian sediments (the Domanik suite), which frames the occurrence area of the Cheslav and Sargaev sediments, consist of dark-gray clayey and bituminous limestones and clayey schists. They are overlain by lagoonal and littoral-marine ferruginated and pyritized sandstones, siltstones, and argillites with thin lenslike interlayers of variegated oolitic allites of the Beidin subzone. In small areas of the southwestern pericline of the Porozhen structure made up of Domanik sediments, the Orlov suite is represented by red bauxites of the Novoprstan facies subzone.

The sediments of the Mendym suite in the western wing of the Suleimanov brachyanticline consist of organogenic-clastic limestones and are overlain by poor-quality variegated oolitic bauxites, allites with ferruginous pellets, and sandstones of the Orlov suite in the western part of the Vyazov facies subzone and partially, in the northern wing of the brachyanticline by mainly terrigenous rocks of the Suleimanov subzone.

The Orlov suite displays the largest facies variability in areas where it occurs on organogenic-clastic limestone of the Upper Frasnian Samsonov suite.

In western parts (near Karatau) organogenic-clastic partly reefogenic limestones of the Samsonov suite are overlain by terrigenous rocks of the Orlov suite, mainly of sand granulometry.

East of the Suleimanov brachyanticline, the distribution of the Orlov suite is more complicated. Samsonov Nimestones there are overlain by sediments of several facies subzones. The Novoprstan subzone is represented by red and gray bauxite and allite, overlain by terrigenous and clay-carbonate rocks. The Vyazov subzone consists of variegated oolitic bauxite, allite, and siallite with ferruginous pellets, oolites, and terrigenous rocks. The Suleimanov subzone includes frequent ferruginous pellet-oolitic allites and siallites, conglomerates, gritstones, and sandstones.

Commercial deposits of SUBD confined to the Novoprstan strip of the Novoprstan subzone tend spatially to be associated with Samsonov limestones of facies zone 2 (see Fig. 2) - the shallow-sea neritic subzone. Bauxites of small deposits in the Kuksha strip of the Novoprstan subzone occur mainly in limestones of facies zone 3 - shallow-sea neritic subzone with intense hydrodynamics. Limestones of the shallow littoral zone with traces of tidal activity (zone 1) are overlain by small bauxite bodies wedging out to the west. They belong to the Novoprstan subzone. The limestones are also overlain by low-quality oolitic variegated bauxites and partly aluminoferruginous and terrigenous rocks of the Vyazov and Suleimanov facies subzones. Thus, in the west of SUBD the boundary separating the facies subzones (Novoprstan and Vyazov) of the Orlov suite effectively coincides with the boundary between facies zones 1 and 2 of the subjacent Samsonov limestones. In the west of SUBD the boundary between the strips (Novoprstan and Kuksha) is close to the boundary of facies zones 2 and 3.

The data suggest that, in occurrence areas of the Late Frasnian continental discontinuity, the boundary of the facies zones (or subzones) of the Orlov bauxite-bearing horizon correspond to the facies and age boundaries of subjacent rocks. This is associated with the inherited nature of the tectonic regime in certain structures. Each facies zone of the Orlov suite is associated with a specific subjacent complex exclusively characteristic of that zone.

The fact that pure Samsonov limestone made up of unequal fragments with different lixiviation susceptibility are karst-forming rocks in certain paleogeomorphological and paleostructural environments is extremely important. On the other hand, the light Samsonov organogenic-clastic limestones of the Karatau area, represented by reefogenic facies, up to 10 m thick, with stromatopore and coral fauna, are overlain exclusively by terrigenous rocks of the Orlov suite, due to the distant location of this region from the source areas of bauxite material. We can say that the necessary condition of bauxite accumulation on limestones of the Samsonov suite is the favorable effect of paleomorphological factors (type of prebauxite topography) and facies factors (distribution of bauxite material sources).

Specifics of Prebauxite Topography

Until recently, there have been virtually no studies of paleotopography formed in Samsonov limestones prior to the onset of bauxite accumulation in SUBD. We attempted a reconstruction of this topography. For the zero surface, we took the practically horizontal (during the Late Frasnian) surface of the base of bituminous limestones of the Ust'-Katav

suite, overlying the Orlov suite. This was the only possible choice, because the nearly 500-m-thick strata of homogeneous limestones of the Ust'-Kataev suite and the overlying strata of facies-similar Famennian limestones, up to 550 m thick, are devoid of reference marks. In addition, the territory is relatively small (15 × 30 km). As a result, the paleotopography was deprived of contrastivity; the lenslike and lenslike-stratal bodies of red diaspor-boehmite bauxites in all instances filled the negative forms of the prebauxite relief. The reconstruction disregarded the thickness of the Samsonov suite and the relationship of thickness and facies composition of the Samsonov and Orlov suites. The erosion-karst or inherited origins of the topography were not discussed.

The thickness of the Samsonov suite in SUBD, established mainly by drilling data, varies abruptly from zero or a few meters in the northeast to 40 m in the southwest of the territory. The averaged thickness in the central area is some 10 m, but there are sharp gradients (over a distance of 2.5 km, the thickness can vary from 40 to 5 m). An analysis of paleoprofiles constructed with adjustment for Samsonov suite thickness (Fig. 3a) determined on isolated areas an inverse correlation between red bauxite thickness and Samsonov suite rocks. Red bauxites were not normally deposited on areas where the suite was thicker than 20 m.

This regularity is typical only for red bauxites. The thickness of variegated low-quality bauxites, gray puritized bauxites, and terrigenous rocks of the Orlov suite overlying the low-quality bauxites and the red bauxites does not correlate with Samsonov suite thickness. A reduction of Samsonov suite thickness can be utilized as a predictive criterion in prospecting for red bauxites.

The initial thickness of the Samsonov carbonate body was largely consistent and even slightly increased in the eastern and northeastern directions. In Samsonov limestones that had been preserved by the deposition time of the Orlov suite, a general thickness reduction was already observed in the contemporaneous structure in these directions. This fact and the nearly ubiquitous presence of bauxitic breccia in the tops of Samsonov limestones under red bauxites indicates that any reduction of Samsonov suite thickness in local areas was due to erosional and erosional-karst processes.

In order to investigate the role of erosional-karst processes, we constructed an additional series of paleoprofiles, taking as the benchmark the base of Samsonov suite limestones overlying clay-carbonate Domanik rocks (see Fig. 3b). The original subhorizontal occurrence of this surface could have been slightly distorted by synsedimentary and tectonic processes during and after limestone accumulation. However, the amplitude of the paleorelief formed by erosional-karst processes appears to be much greater. It is unlikely that within a single facies zone of a shallow-water shelf, abrupt variations of limestone thickness (up to 20-40 m) over a length of just 2-3 km could be caused entirely by synsedimentary processes. The paleoprofiles exhibit a series of erosional-karst lows and valleys in the paleorelief, divided by watersheds. Red bauxites are associated with negative forms of this paleorelief and tend to occur closer to the axial and central parts and less often in slopes. The paleorelief does not exert any significant influence on the distribution of other rocks of the Orlov suite (variegated and gray layered bauxites and terrigenous rocks).

An erosion of the Domanik and Samsonov suites with a total thickness of 70 m in the northeast of the territory (Porozhen paleoupland) prior to the onset of bauxite accumulation, and possibly during its early stages, indicates a general low-angle slope of the prebauxite topography surface in the southwestern direction toward the Uluir paleodepression.

West of the Uluir paleodepression, another positive paleostructure existed during the Orlov time: the Suleimanov paleorise (at the site of the present Suleimanov brachyanticline), which controlled the washdown of terrigenous matter. Judging by the distribution of terrigenous facies of the Orlov horizon, the Suleimanov upland during the Orlov time was more contrastive than the Bashkirian upland arch, situated to the north.

Two strips of continental red bauxites stretching in a southwesterly direction (to the dip of the Porozhen paleorise slope) - Novoprystan and Kuksha - correspond to broad valley-like erosional lows in the relief, worked out in Samsonov limestones. The longest Novoprystan strip is traceable for 30 km, with a width from 3 to 6 km. It is connected with the Kuksha strip by a transverse valley, 2-3 km wide and 13-15 km long. The "cut" of these depressions is tens of meters deep. These negative relief forms in carbonate rocks could not be entirely attributed to karst processes. These processes probably evolved before the poljes

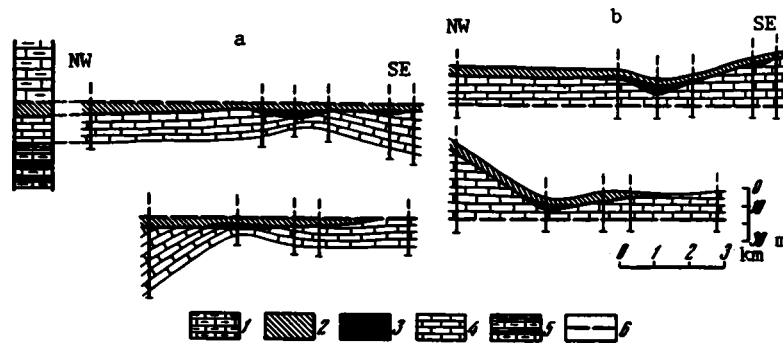


Fig. 3. Lithologic-facies profile of the Orlov and Samsonov suites in the SUBD area at the base of the Ust'-Katav (a) and Samsonov (b) suites: 1) gray and dark-gray clayey limestones, Ust'-Katav suite, Upper Frasnian; 2-3) Orlov suite, Upper Frasnian (2 - diaspora-boehmitic bauxites, 3 - gray bauxites, variegated oolitic bauxites, allites, siallites, argillites, siltstones, and sandstones); 4) light-gray and gray massive finely crystalline limestone, Samsonov suite, Upper Frasnian; 5) dark-gray clayey limestone, black argillite, Domanik suite, Lower Frasnian; 6) paleoprofile reference benchmark.

connected by tops were formed here; they were then succeeded by erosional processes. Commercial deposits of SUBD are located in the Novoprstan strip; the Kuksha strip contains smaller placers of high-quality bauxites of the Kuksha group. The Ailin strip of the same orientation is the shortest and the shallowest. It corresponds to the northwesternmost "valley." Small bauxite shows of high and medium quality are associated with this strip.

Erosional-karst processes are developed most intensely on weakened areas, usually confined to fissured zones, controlled by large surrounding faults. These faults in the SUBD area occur in the long Novoprstan northeasterly downthrow, located to the west of the Novoprstan strip, and a system of northeasterly faults belonging to the zone of the regional Berdyaush overthrust, which coincides here with the eastern boundary of the Chuskovskaya zone.

To sum up, the study of the age, composition, lithologic and facies zonality, and thickness of the deposits underlying the bauxite-bearing horizon has determined several bauxite accumulation factors that can be utilized to develop various predictive criteria.

1. The existence of a zone of carbonate bodies and Samsonov-type structures in the margins of the Chusovskaya zone. They are typically bounded by contemporaneous relatively deep-sea (Mendym and Domanik type) or lagoon (Mendym) sediments. These carbonate bodies were developed on positive paleostructures of ancient formations.

2. Slightly sloping plateaus of Samsonov limestones deposited in margins of shallow-sea shelves and shallow-water areas with intense hydrodynamic and neritic zones with active or slow water movement provided most favorable beds for bauxite formation.

3. The development of paleorises during the Orlov time with low-angle slopes made up of Samsonov limestones. Some of the paleorises (such as the Porozhen upland) developed in the margins of the Chusovskaya zone, where the accumulation area of bauxitic material are situated.

4. The largest thickness of red bauxite beds is found in areas with minimal thickness of the Samsonov suite; this is associated with intensity of erosional and erosional-karst processes (see Fig. 2).

5. The existence of isolated bore holes with bauxites in favorable areas determined on the basis of the underlying rock complex makes it possible to predict erosional-karst depressions with bauxite placers typical for SUBD.

LITERATURE CITED

1. L. A. Antonenko, M. A. Beér, and A. A. Byvshev, "Geological principles of a detailed forecasting of the Upper Devonian bauxites in South Ural," in: Large-Scale and Local Forecasting of Aluminum Deposits: Principles, Evaluation, and Methods [in Russian], VIMS, Moscow, pp. 58-68 (1986).
2. N. I. Arkhangel'skii, "Occurrence settings and genesis of Ural bauxites," in: Proceedings of a Conference on the Genesis of Iron, Manganese, and Aluminum Ores [in Russian], Izd-vo Akad. Nauk SSSR, Moscow, pp. 12-23 (1937).
3. M. A. Beér and A. A. Byvshev, "Main stages in the development of bauxite-bearing structures of the Devonian shelf in West Ural," in: Forecasting and Evaluation of Bauxite Deposits [in Russian], VIMS, Moscow, pp. 33-45 (1985).
4. B. M. Mikhailov (ed.), Bauxite-Bearing Rock Complexes in the Ural [in Russian], Nedra, Leningrad (1987).
5. A. A. Burlakov, E. V. Ershova, A. V. Leiptsig, et al., "Distribution and structure of a bauxite bed in premineralization topography of deposits in the Petropavlovsk subzone, North Ural," in: Forecasting and Evaluation of Bauxite Deposits [in Russian], VIMS, Moscow (1983), pp. 82-92.
6. S. M. Domrachev, "The Devonian of the Karatau ridge and adjacent South Ural areas," in: The Devonian of the West Ural Region [in Russian], Gostoptekhizdat, Moscow (1951), pp. 2-119.
7. A. V. Leiptsig, V. N. Vil'shanskii, and Yu. E. Kustov, "Paleotopography and localization features of paleogene bauxites in the Amangel'dy mineralization district," in: Forecasting and Evaluation of Bauxite Deposits [in Russian], VIMS, Moscow (1983), pp. 116-126.
8. S. V. Maksimova, Ecologic-Facies Features and Formation Conditions of the Domanik [in Russian], Nauka, Moscow (1970).
9. O. M. Mkrtchyan, Distribution Patterns of Structural Forms in the Eastern Russian Platform [in Russian], Nauka, Moscow (1980).
10. A Handbook on Lithology [in Russian], Nedra, Moscow (1983).
11. N. M. Strakhov, "Domanik facies of the South Ural region," Tr. GIN. Geol. Ser., Issue 16, No. 6 (1930).
12. A. P. Tyazheva, "Stratigraphy of the Devonian from the western slope of the South Ural," in: The Devonian of Bashkiria [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1961), pp. 5-128.
13. J. L. Wilson, Carbonate Formations in Geologic History [Russian translation], Nedra, Moscow (1980).

ALTERATION OF THE MICROTEXTURE OF ORGANICS
DURING CATAGENETIC EVOLUTION

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UDC 553.983:552.14

The alterations of microtextures of various coal microcomponents and dispersed organic matter are studied in correlation with the stage of catagenetic maturation. The microtexture of certain components of organic matter can be used to judge about the degree of thermal evolution, particularly the main oil formation phase.

Organic matter (OM) is the sedimentary rock components most susceptible to external thermodynamic factors. It is extremely sensitive to variations of the physical and chemical parameters of the environment. During formation and subsequent lithification of sediment, the initial OM experiences most alteration in two phases. First, during the early phase of sediment formation in subsurface conditions, OM is subjected to a greater or lesser oxidation, depending on the burial conditions. Secondly, after the subsiding sediments reach a certain depth and a temperature T_{f0} , OM begins to generate bitumoids with a high intensity. This phase is described as the main oil formation phase (MOFP) in OM catagenetic evolution.

During the first (oxidation) and second (oil and gas generation) phases, the OM volume becomes reduced. This produces a secondary porosity V_{BT} [1], resulting in a microtexture formed in the initial OM. Thus, OM microtexture depends, first, on its primary chemical composition and degree of oxidation during sedimentation and, secondly, on the degree of catagenesis.

Assuming that the different microcomponents making up OM are altered differently, one should expect that V_{BT} may also vary; each microcomponent of the microtexture observed on the broken specimen surface will have its specific value of V_{BT} .

We investigated OM microtexture under a JSM-50A raster electron microscope (REM). The study was conducted with secondary electrons with magnifications of 1000-10,000. When dielectric materials are investigated in an REM, a layer of conducting material is coated on the specimen surface. In order to avoid a distortion of surface morphology all specimens were prepared as follows: a gold film of 300 Å was sprayed in a vacuum in a Fine Coat chamber. Gold film of this thickness produces an optimal secondary-electron emission without distorting fine surface details [5].

To prevent ambiguities in the interpretation of the resulting photographs, we began by studying concentrated OM forms - coals. Coal specimens with diverse microcomponent compositions were collected from depths from 2 to 3 km and T_{f0} from 70 to 110°C, respectively (see Table 1).

The microcomponent composition of the coals was determined on transparent polished sections. Based on the coal petrography, three groups of specimens were prepared, or areas were identified in coals; in group 1 these were mainly fusinite, in group 2 vitrinite, and in group 3 liptinite.

Remarkably, coal microlayers of 1-3 mm, selected from Bazhenov suite deposits in various locations (see Table 1), were made up of virtually the same vitrinite. The other coal specimens contained substantial quantities of each of the three components.

The degree of catagenetic alteration of OM in the coal specimens increased regularly with increasing T_{f0} . This is confirmed by derivatographic studies (volatile yield per organic weight and increase in the maximum volatile yield temperature) [3]; it was also reflected in the variation of OM pigmentation in transparent polished sections (darkening to

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TABLE 1. Microcomponent Composition and Selected Physical Properties of Coals from Shirotnoye Priob'ye Deposits

Deposit	Well number	Sampling depth, m	Formation temp., °C	Microcomponent composition, %			Volatile yield, %	Max. volatile yield temp., °C
				fusinite	vitritinite	lip-tinite		
Shushma	1081	1993	70	92	5	3	21	430
Solkin	1152	2742	82	95	3	2	16	435
Teplov	274	2888	91	97	2	1	23	440
Talin	3902	2740	90	75	20	5	17	440
Var'egan	2153	2618	80	3	95	2	40	390
Tyumen	121	2756	85	0	100	0	32	430
Salym	131	2918	110	0	100	0	20	460
Northern Var'egan	4P	3130	102	5	70	25	40	435

dark-brown and almost black in specimens collected from the deepest bore holes with maximum T_{fo}).

After a detailed investigation of broken surface microtexture morphology in concentrated OM forms (coals), the surface textures of scattered OM (SOM) were also studied. For this study we took rock specimens rich in SOM - black schists (kukersite in the Bazhenov and Mennilitov suites).

DISCUSSION

The effect of oxidation on initial OM microtexture can be traced conveniently in fusinite, because it is most oxidized coal microcomponent, with carbonatization and blackening. By contrast, catagenetic alteration of OM microtexture should be clearly seen during thermal maturation of vitrinite and liptinite components, which were much less oxidized than fusinite in the early stages of sedimentation.

By means of simple balance calculations based on elemental composition of coals with various sets of microcomponents, bitumoids, and nonhydrocarbon volatiles and on published data [4, 6], one can estimate approximately the maximum generation properties G_{max} of various coal microcomponents. G_{max} is defined as the active part of OM, i.e., the part that is converted to bitumoids and nonhydrocarbon volatiles during the course of catagenetic evolution.

Thus, G_{max} in fusinite varies from 10 to 25% of the original weight, depending on the coal properties and composition. In vitrinite, G_{max} attains some 40-50% and in liptinite as high as 60-70% of the initial weight or volume. In absence of compactification, generation processes can result in V_{BT} changing from 10 to 70%, depending on microcomponent composition.

Based on these estimates, one should expect that minimum textural alteration of OM_{ini} surface during the course of thermal maturation would be observed in fusinite. Vitrinite should display a more noticeable change in microtexture; the liptinite component of OM_{ini} should have the largest degree of alteration. The morphology of the microtexture of various OM_{ini} microcomponents produced by generation of bitumoids and nonhydrocarbon volatiles is also affected by several other factors. First, it is influenced by the initial texture of microcomponents; secondly, by the geometric shape of their presence in the rock (spherical bodies, films, crustlets, or massive formations); thirdly, by the strength of their bonds with the mineral matrix; and, fourthly, by the rheologic properties and capacity for withstanding the lithostatic pressure.

Phyterals are a simple example of the influence exerted by the internal texture of microcomponents on the morphology of the resulting V_{BT} and its microtexture. This is due to their anisotropic build and the resulting variations of V_{BT} formed after different components of phyterals (Fig. 1c).

The effect of initial OM upon its microtextural morphology can be illustrated by spherical, film-shaped, and massive formations. Oxidation and generation of bitumoids and nonhydrocarbon volatiles from OM are processes that are physically similar to dehydration.

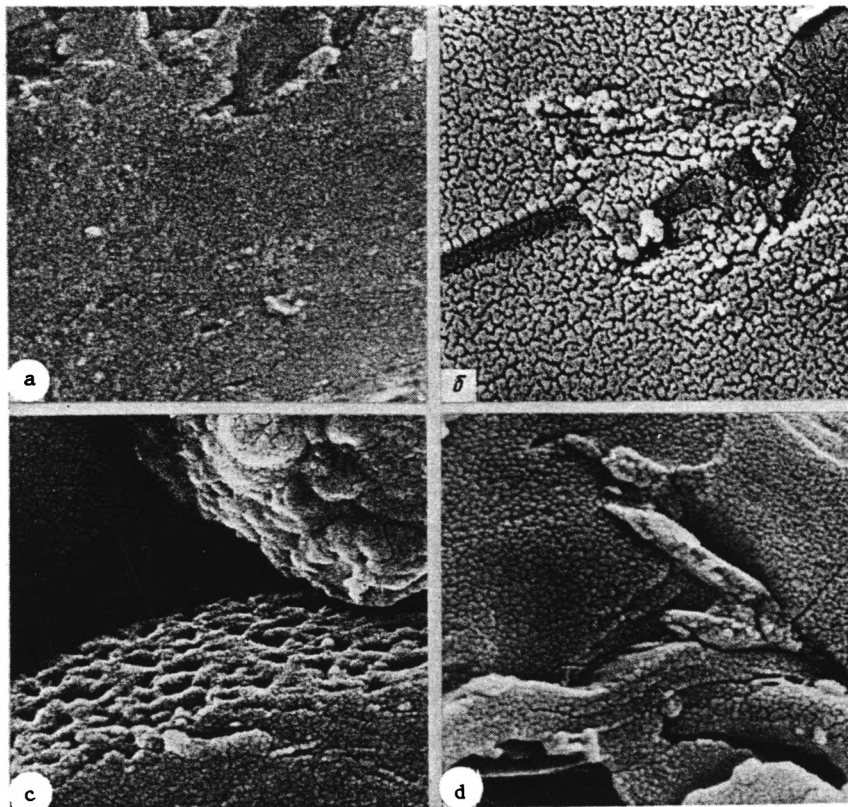


Fig. 1. Microtextures of broken surfaces of fusinitic coal microcomponents. Deposits: a) Shushma, b) Solkin, c) Teplov (a phyteral fragment at bottom), d) Talin. Magnification $\times 10,000$.

Accordingly, the morphology of V_{BT} or OM microtexture can be described by methods employed in studies of dehydration processes [2].

Specifically, for spherical OM bodies subjected to oxidation or generation of bitumoids and volatiles, three main routes of V_{BT} formation are possible.

If the OM_{ini} is amorphous with a perfectly isotropic texture, the oil generation process will "compress" the initial sphere, reducing its radius. However, OM is usually texturally inhomogeneous, with two other alternatives of V_{BT} formation.

If a spherical OM body is textured so that generation of bitumoids and volatiles results mainly in its reduction along the radius, cracks will be formed, opening into the interior of the organic mass.

If generation processes mainly shrink OM perpendicular to the radius, the cracks will open on the outside.

In both cases, the initial radius of a spherical OM body will not change. When OM exists in the form of thin films, microlenslets, etc., a texture will be formed similar to a dried-out clay crust, consisting of polygons separated by "desiccation" cracks.

The action of the third order – the degree of bonding of OM_{ini} with mineral substrate – can be illustrated by the shapes of cross-sections of the "desiccation" cracks. Two alternatives are possible. If the bonding of OM with the substrate is weak, the cracks separating individual organic microblocks have vertical boundaries. When OM_{ini} is strongly bonded to the substrate, the crack cross-sections are conical, expanding toward the OM free surface.

In massive OM_{ini} formations the network of microcracks is similar to that in films, but the network is three-dimensional (rather than two-dimensional), dividing the organic mass in micropolyhedra. The three-dimensional network is not closed completely, some cracks are not connected, the polyhedra retain some bonding and do not fall apart.

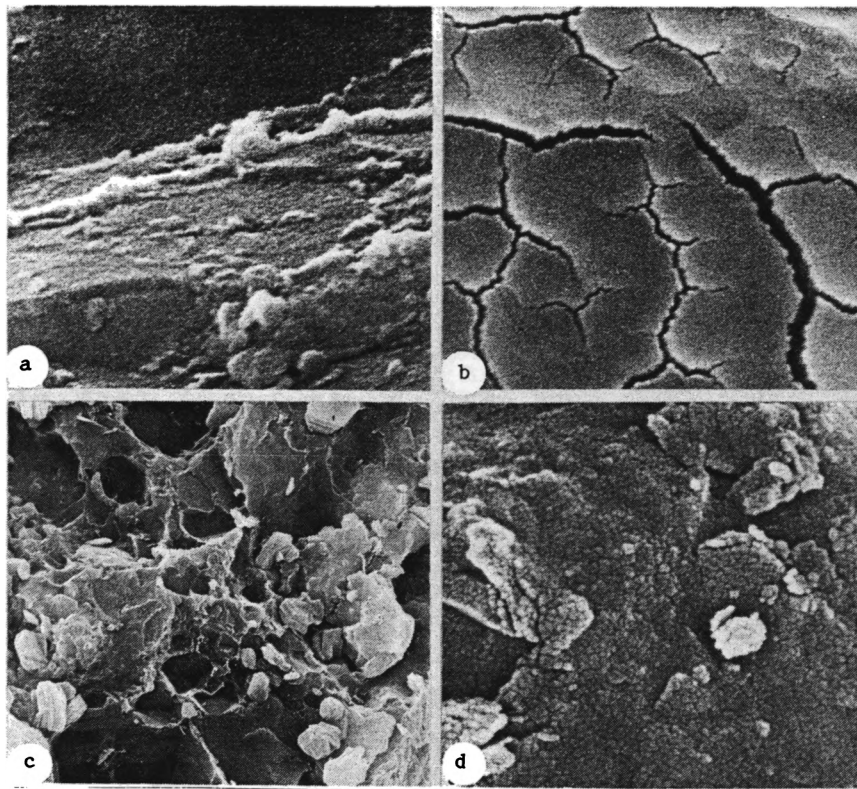


Fig. 2. Microtexture of broken surfaces of vitrinitic microcomponents of coals from the Bazhenov suite. Deposits: a) Var'egan, b) Tyumen, c) Salym, d) Salym (enlarged fragment of c). Magnifications: a, b, d - $\times 10,000$, c - $\times 3000$.

The fourth factor is the rheologic and strength properties of OM types. It is responsible for preservation or disappearance of V_{BT} in the form of microcracks as organic mass "flow" into the cracks under the effect of surface tension or lithostatic pressure.

Microtextures of oxidized varieties of coal microcomponent (fusinite) were studied on specimens at different T_{fO} (see Table 1). The coals are mainly of a fusinitic composition, with unequal volatile yields. The maximum volatile yield was observed in the temperature interval of 430-440°C (see Table 1). There was no correlation between the microcomponent composition of the coals, the volatile yield, and T_{fO} . However, there was some positive correlation between T_{fO} and the maximum volatile yield temperature T_{max} (see Table 1).

On REM photographs, regardless of T_{fO} and volatile yield, the broken surfaces of all fusinite specimens had a distinct microtexture. It was formed of micropolyhedra or microblocks with transverse dimensions of 0.1-0.4 μm , separated by cracks with openings of 0.01-0.08 μm (see Fig. 1). The lowest-temperature specimen from Shushma deposit had a less distinct fusinite microtexture than did other specimens with higher T_{fO} (Fig. 1a). Subsequent catagenetic alteration was likely to intensify the relief of the microtexture formed in fusinites. It is also possible, however, that the Shushma specimen was less oxidized during sedimentogenesis and diagenesis than were fusinites from other deposits.

Importantly, microblocks formed by oxidation in fusinites are comparable in size to the light wavelength (violet spectral region). Since the size of microblocks (micropolyhedra) varies little within a fusinite specimen, the microtexture imparts the properties of a diffraction grating to the broken coal surface. Incident light interacts with the surface and endows it with a peculiar silky tinge.

It appears in coals with relatively large microblocks (near 0.4 μm or more); in smaller polyhedra, diffraction affects ultraviolet rays invisible to the human eye.

The textural features of fusinites do not provide clear information on their catagenetic alteration. The microtexture is formed mainly by oxidation of initial vegetable tissues and

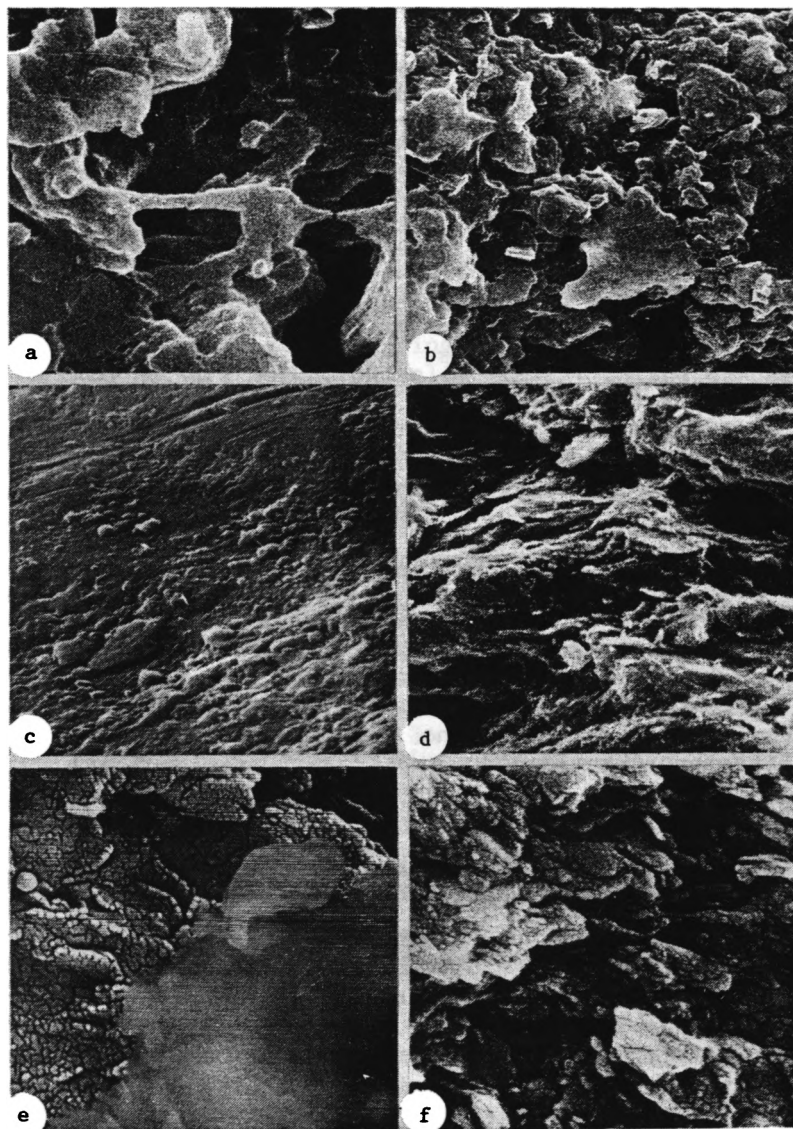


Fig. 3. Microtexture of broken surfaces of liptinite bitumens and SOM of black shales. Deposits: a) Northern Var'egan (liptinite); b) Northern Var'egan (metamorphosed vein bitumen); c) Salym (unmetamorphosed cracked bitumen); d) Oktyabr'skoe (kukersite); e) Kholmogory (Bazhenov suite); f) Orlov (menilite suite).

is affected secondarily by subsequent natural thermal destruction in MOFP.

A convenient object for analysis of the alteration of microtexture of OM concentrated forms during catagenetic processes is provided by coaly microlayers from bituminous sediments of the Bazhenov suite, effectively all consisting of a single microcomponent: vitrinite (see Table 1).

The Var'egan specimen was found at T_{f0} of 80°C. Its catagenetic alteration was minimal among all Bazhenov coals. The coal in this specimen had the maximum (40%) volatile yield and minimal T_{max} (see Table 1). Its broken surface microtexture is even and uniform (see Fig. 2a).

Bazhenov coal from Tyumen deposit was slightly more altered. It existed in the bed at $T_{f0} = 85^\circ\text{C}$. The volatile yield from this coal was 32%; $T_{max} = 430^\circ\text{C}$ (see Table 1). A network of desiccation microcracks appeared on the broken surface, with a crack opening of $\sim 0.2 \mu\text{m}$. The crack network divided the broken surface into microblocks of 1-2 μm (see Fig. 2b).

Specimens from the Bazhenov suite of the Salym deposit exhibited the highest degree of catagenetic alteration. The present T_{f0} of these rocks is 110°C. The volatile yield is just 20%; $T_{max} = 470^\circ\text{C}$ (see Table 1).

An analysis of REM photographs determined a complex structure of the broken surfaces of coal specimens from the Saly deposit. The surface displayed numerous microcaverns of 2-8 μm , often containing pseudo-hexagonal authigenic kaolinite crystals (see Fig. 2c). Like fusinite, the organic mass acquired a microcellular or microblock fabric, with the average linear microcell polyhedra of 0.4 μm and crack opening of 0.01 μm (see Fig. 2d).

The microcavernous and microcellular textures of the coal are at least partly responsible for an abrupt increase in open porosity in Bazhenov suite specimens from high-temperature areas of the Saly deposit (as high as 6-8%). The organic mass itself has a much higher porosity.

Microtextures of vitrinite coal specimens collected from bituminous Bazhenov suite rocks indicate that intensive generation of bitumoids and volatile products, creating V_{BT} and corresponding microtexture of the organic mass, began at $T_{fo} > 80^\circ\text{C}$.

The upper boundary of MOFP for coals with this composition lies at the level of isotherms of 80-85°C. We should mention, however, that, according to some authors, present T_{fo} are much lower than those that existed previously (according to various estimates, paleotemperatures were higher than modern levels by 20-50°C). That would probably raise the upper MOFP boundary to 100-130°C.

The microtextures of liptinite microcomponents showed that in specimens at relatively high T_{fo} (about 100°C) there was no microblocks typical for fusinites and heavily catagenetically altered vitrinites. However, there was a "lace" texture formed of numerous micropores and microcaverns with an average size of 1-3 μm in the liptinite mass (Fig. 3a). Presumably, the liptinite mass preserves sufficient plasticity even in such harsh thermodynamic conditions. As a result, surface tension forces formed in a liptinite body (after bitumoid generation) heal the microcracks. The liptinite mass becomes compressed, and microcavities observed in photographs are formed.

It is possible that during a later, more advanced catagenetic stage, as liptinite exhausted its generation potential (the volatile yield was reduced), the rheologic properties of liptinite became changed. Microcracks were no longer healed, and a microblock microtexture was formed similar to fusinites or heavily catagenetically altered vitrinites.

Viscous bitumens are similar in properties and composition to liptinite. The surface microtexture of heavily metamorphosed vein bitumen from the northern part of the Var'egan deposit (a depth of 3496 m) was similar to liptinite (see Fig. 3b). Like liptinite, the metamorphosed bitumen develops a substantial V_{BT} after volatile components are released from it at high T_{fo} . Microstrips of 1-3 μm and larger are formed. Like in liptinite, the metamorphosed bitumen has no microblock microtexture (see Fig. 3a, b), which may develop at a later, more advanced stage of catagenesis. Ordinary nonmetamorphosed bitumen exhibits a smooth broken surface similar to that of vitrinite or liptinite of a low catagenesis stage, preceding MOFP (see Fig. 3c).

Microtextures of broken surfaces of rocks enriched in OM (SOM) and often described as black or bituminous shales (e.g., Bazhenov and Menilitov suites and kukersite) suggested an alteration of films or crustlets of SOM sorbed by mineral particles during thermal maturation similar to the earlier-described microtextures of broken surfaces of vitrinite coal [7].

A virtually unaltered SOM of kukersite, which, despite its old age, has never participated in MOFP, has an even surface, like vitrinite from the Var'egan area (see Figs. 2a and 3d). Specimens collected from the Bazhenov and Menilitov suites participated in MOFP. Accordingly, the SOM in these specimens generated bitumoids and volatiles. As a result, V_{BT} and the corresponding microtexture are similar to those of vitrinites of advanced catagenesis (see Figs. 2d and 3e, f) or fusinite (see Fig. 1b-d). SOM is distinguished from OM of coals in that in black shale the mineral matrix texture is "transparent"; it is completely suppressed in coal surface microtexture.

In an analysis of SOM microtexture and coal microtexture, we differentiate between mature sediments that participate in MOFP and "immature" sediments found above MOFP; the distinctive feature is the presence or absence of a mosaic microblock texture of the SOM surface. The possibility of an oxidative setting in marine sediments enriched in SOM cannot be ruled out completely. In these settings SOM may be oxidized and form a microblock texture similar to that of fusinite, which in this case would characterize the oxidation degree rather than the catagenesis stage. Such oxidized SOM could only be differentiated from catagenetically altered matter in a special bituminologic investigation and by T_{max} .

The following conclusions have been drawn from this study:

1. Oxidation (fusinization) and thermal evolution during catagenesis produce a mosaic microblock surface of the initial OM, presumably due to a reduction of OM volume as the oil generation potential becomes used up.
2. A similar surface microtexture is observed in fusinite coal components in the entire range of T_{fo} studied (from 70 to 100°C): vitrinitic coal at $T_{fo} > 85^{\circ}\text{C}$. It is absent in liptinitic coals.
3. OM microtexture developing as a result of formation of V_{BT} is affected not only by the microcomponent set and the degree of catagenetic alteration but also by the shape of microcomponents and the strength of bonds with the mineral matrix.
4. SOM and concentrated forms (coals) during the course of catagenetic evolution acquired a microblock surface texture morphologically similar to the cracking texture of a clay crust.
5. The presence of such a SOM microtexture indicates that the rock has entered MOFP.
6. Vitrinite is the most suitable microcomponent for evaluating the degree of catagenesis on the basis of OM microtexture.

LITERATURE CITED

1. M. Yu. Zubkov, V. A. Ershov, I. A. Pryamonosova, and A. Kh. Shakirova, "Generation of bitumoids and formation of capacity space in Bazhenov suite rocks," in: Progress in Exploration and Development of Oil and Gas Deposits in Western Siberia [in Russian], ZapSibNIGNI, Tyumen (1984), pp. 16-21.
2. M. Yu. Zubkov, "Nodule formation models," in: Postsedimentary Mineralization in Sedimentary Formations [in Russian], ZapSibNIGNI, Tyumen (1985), pp. 43-51.
3. M. Yu. Zubkov, T. A. Fedorova, and I. A. Pryamonosova, "Alteration of microtexture of organic matter in the Bazhenov suite during thermal maturation," in: Proceedings of the Fifth Annual Conference on the Geology and Mineral Resources of the West Siberian Plate and Its Surrounding Folded Structures [in Russian], ZapSibNIGNI, Tyumen (1987), pp. 222-223.
4. A. E. Kontorovich, I. I. Nesterov, F. K. Salmanov, et al., Oil and Gas Geology in West Siberia [in Russian], Nedra, Moscow (1975).
5. J. Gouldstwein and H. Jacowitz, Practical Raster Electron Microscopy [Russian translation], Mir, Moscow (1978).
6. B. Thissault and D. Welte, Oil Formation and Spread [Russian translation], Mir, Moscow (1981).
7. T. A. Fedorova, "Alteration of microtexture of adsorbed organic matter in clay oil-parent rock during catagenesis," in: Proceedings of the Fourteenth Conference of Young Geologists at Moscow University Geology Department [in Russian], Manuscript deposited with VINITI, No. 7154, Moscow (1987), pp. 162-167.

EXPERIMENTAL STUDY OF THE REDUCTION DEPOSITION OF URANIUM BY PYRITE UNDER STANDARD CONDITIONS

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UDC 550.4:553.495

The role of iron disulfide, pyrite, as a possible reducing agent for uranium in rocks having no microbiologically active organic matter is discussed. The electrode potential of freshly exposed pyrite surfaces of various origins was determined experimentally, and the ability of this surface to reduce U(VI) and cause it to precipitate out of practically oxygen-free solutions was demonstrated.

It is known that the uranium in uraniferous rocks is ordinarily associated with pyrite. Despite the number of works dealing with the role of pyrite in a uranium deposition, there is as yet no single accepted explanation of it.

The apparatus of chemical thermodynamics is being used with increasing frequency to provide quantitative descriptions of natural processes. Since the rate at which equilibrium is attained is not a factor in thermodynamic calculations, and since reactions are often very slow under supergene conditions, and some of the thermodynamic requirements are not even satisfied, it is important to determine exactly which equilibria exist in nature.

If the system $S_2^{2-} + 8H_2O \rightleftharpoons 2SO_4^{2-} + 16H^+ + 14e^-$ is considered to be reversible, then calculations show that this system lies below the equilibrium of uranyl-carbonate and uranyl-sulfide complexes with uranium oxides of the series $UO_2-UO_{2.25}$. In this case, pyrite must reduce uranium. This clearly follows from the phase diagrams of R. M. Garrels and C. M. Christ [4] and A. A. Abramov [1], and from the calculations of R. P. Rafal'skii [10], who used the method of I. K. Karpov [6], and from calculations made by Shade Markus [14].

According to the data of S. S. Zavodnov [5] and M. F. Stashuk [13], the systems $S^{2-} + 4H_2O \rightleftharpoons SO_4^{2-} + 8H^+ + 8e^-$ cannot be completely reversible, so that the potential-determining system under exogenic conditions is considered to be the system $S^{2-} \rightleftharpoons S_{cr} + 2e^-$. Because of this, A. K. Lisitsin [8] and E. M. Shmariovich [15] based all their calculations on the equation for the second reaction. In this case the equilibrium line for the system $S_2^{2-} \rightleftharpoons 2S_{cr} + 2e^-$ lies above the Eh-pH field for the reduction and deposition of uranium from stratal water, from which it follows that pyrite cannot be the reducing agent for uranium.

It should be pointed out that the upper boundary of the pyrite stability field calculated by the two authors was in each case confirmed experimentally [2, 8].

The disparity between the experimental data, due primarily to using different techniques [2, 8, 9, 12] and possibly to the semiconducting properties of pyrite (conductivity type) [9], may be the source of the differing conclusions reached concerning their origin.

In addition to measuring the electrode potentials of pyrites, workers have also done direct experimental work on the ability of pyrite to cause pyrite to deposit. H. Granger and C. Warren [16] proposed that during the slow oxidation of the pyrite of the host rock, unstable, soluble sulfur compounds such as sulfite and thiosulfite form that slowly decompose, ultimately forming H_2S and SO_4^{2-} , thereby creating the physico-chemical conditions necessary for uranium reduction in a biologically sterile environment. To support their concept, Granger and Warren modelled a process for the formation of sulfites and thiosulfites during the oxidation of freshly precipitated sulfides of the sandstone by a solution of sodium bicarbonate containing oxygen. These workers also succeeded in obtaining pyrite and uranite from sulfite and thiosulfate solutions. It should be pointed out, however, that the proposed

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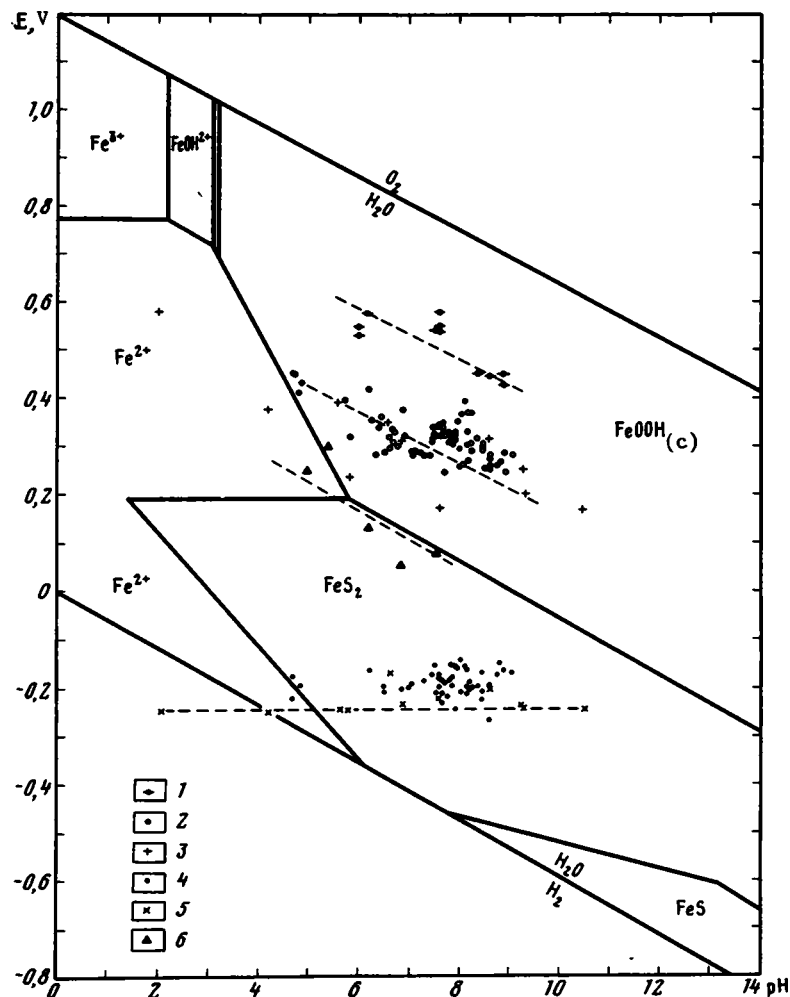


Fig. 1. Results of measurements of the electrode potential of a freshly exposed pyrite surface in solutions bubbled with nitrogen, and values of the Eh of water-saturated (originally oxalic water) pastes with powdered pyrite. 1-3) Eh of the solutions (1 - oxygen-containing, 2 - oxygen-free, in a cell with a blade, 3 - the same, in a cell with no blade); 4, 5) ϕ -potential of a freshly exposed surface (4 - cut with the blade, 5 - cleaved); 6) Eh of water-saturated pastes.

model is far from reflecting natural conditions. First, the thiosulfates were observed to form at pH 9, i.e., under appreciably more alkaline conditions than those in the infiltration-type uranium deposits. Second, one can hardly substitute freshly precipitated Fe(II) sulfide for pyrite, since they have different oxidation-reduction properties.

Another attempt to determine experimentally the role pyrite plays in the deposition of uranium was made by A. V. Kochenov and Yu. L. Medvedev [7]. They used pyrite from concretions of sedimentary rocks. In the near-neutral environment, the Eh was observed to fall to -300 mV, and uranium was precipitated. The experiments were not performed under sterile conditions, however, and the authors themselves admit that microorganisms could have been involved in the process. It appears to us that the mechanism whereby the Eh of the environment was lowered and uranium precipitated in the experiments of A. K. Kochenov and Yu. L. Medved is a bacterial one. Experimental modelling of bacterial precipitation of uranium in gray-colored rocks of the pyrite geochemical type was conducted by E. V. Rozhkova, E. G. Vasil'eva, and E. G. Kuznetsova [3, 11]. It was then shown that the Eh of the environment in unoxidized rocks containing pyrite and organic matter is lowered and that uranium is deposited under the influence of the products of the vital activity of sulfate-reducing, hydrogen-producing, and other types of bacteria.

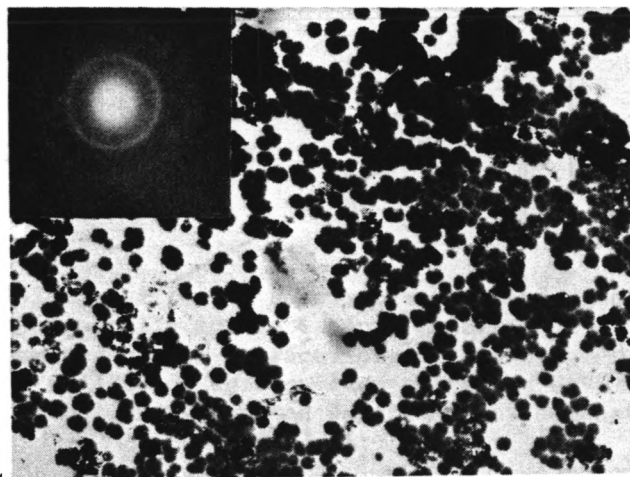


Fig. 2. Orbicular structures of uranium oxide UO_2 on the surface of pyrite (sample BK) after the experiment, replica with extraction, magnification 13,500, and x-ray diffraction images obtained from them.

TABLE 1. Results of Microchemical Determination of a Series of Elements in the Samples of Pyrite Studied, g/ton

Sample number	Co	Ni	As	Pb	Zn	Cu	Mn
29	1480	60	150	80	Not det.	50	60
K2b	160	450	170	10	70	10	70
K-1	100	410	140	110	80	30	70
K-4b	410	500	690	240	110	40	290
BK	180	1920	510	160	130	210	110
66b	60	520	810	230	130	50	60
124	80	220	150	160	70	40	380
125	100	260	Not det.	190	70	20	30
8/479	50	250	170	200	140	20	100
1/2203	200	530	230	190	90	30	50

Analyst, S. P. Pyrusova.

Thus, the role of pyrite in the formation of geochemical barrier reducing barriers and the deposition of uranium in bacterially sterile environment remains unclear.

To answer these questions, the author carried out an experimental study on the oxidation-reduction properties of pyrite and how these relate to the reduction deposition of uranium from bicarbonate waters under bacterially sterile conditions.

The samples selected for the study were of different origins and differed in degree of perfection of their crystal structures, content of impurity elements, and physical and electrical properties (semiconductor type). The semiconductor type of the starting samples was determined by G. A. Gorbato, the mineralogical studies were performed by B. G. Kruglova, and the electron-microscope studies by V. T. Dubinchuk. The samples collected from hydrothermal deposits (the Akchatau tungsten deposit, the Kovdor iron deposit, and the Berezovskii gold deposit) as well as various types of sedimentary rocks (clays, clayey siltstones, and marls) were provided by V. E. Golomolzin, M. F. Maksimova, I. A. Kondrat'eva, G. A. Mashkovtsev, N. V. Skorobogatova, and A. S. Stolyarov.

The oxidation-reduction properties of the pyrites were studied by two different types of experiments: 1) measurement of the electrode potential ϕ of the fresh surface of the pyrite in solutions bubbled with nitrogen, and 2) measurement of Eh-pH equilibrium values of the solution in a water-saturated paste of originally oxidic water and powdered pyrite (for the 0.1-0.5 mm fraction).

The ϕ -potential measurements were made in a cell based on a design of A. A. Abramov [2]. The measuring electrode was a crystal of the pyrite itself, which had a specified portion of its surface in contact with the solution. The reference electrode was a silver-chloride

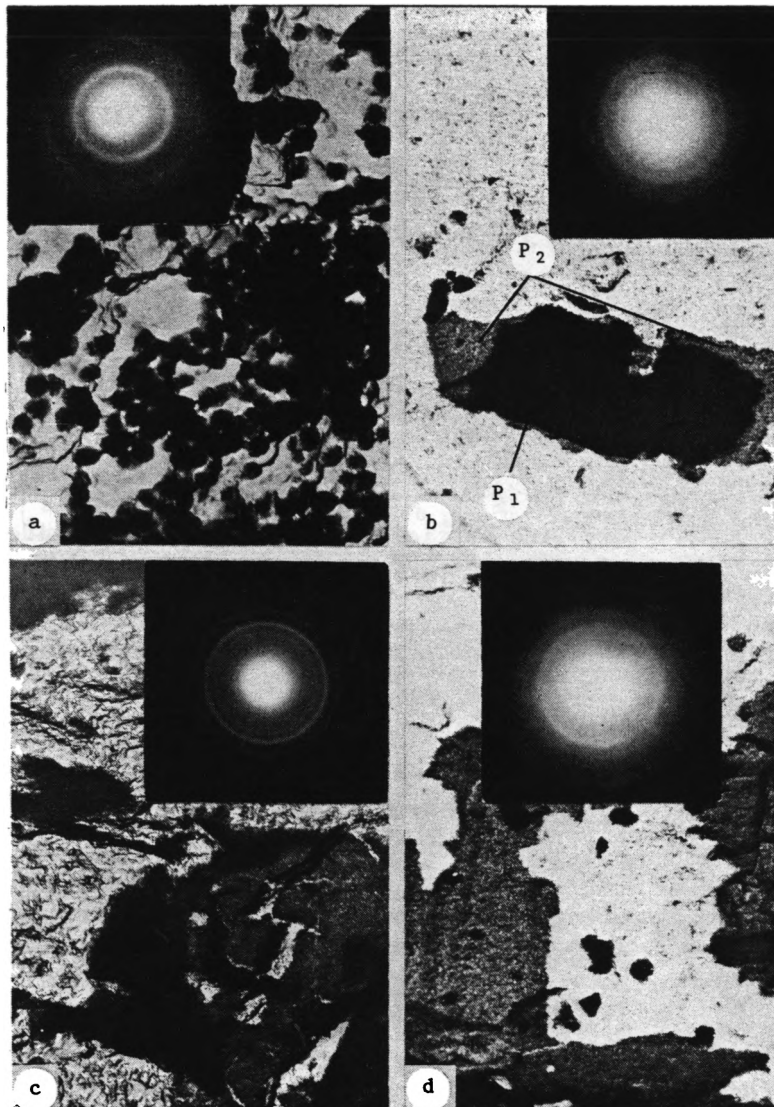


Fig. 3. Orbicular and film structures on the surface of the original pyrite, P₁ (sample 1/2203). a) uranum oxide UO₂, magnification 10,500; b) pyrite (P₂), magnification 7500; c) magnetite, magnification 10,500; d) wustite, magnification 7500, replica with extraction. Insets: microdiffractograms of the structures.

electrode. In addition, the cell was equipped with electrodes to measure solution Eh and pH. To prevent oxidation of the pyrite electrode cut from the sample, it was covered (except for the end) with a layer of organic glass [PMAA] dissolved in chloroform. The design of the cell provided for isolation of the deoxygenated solution from the air, even while the surface of the pyrite was stripped with a corundum blade. The ϕ -potential was measured during the process of exposing the pyrite surface and after the process was completed.

In addition, a second series of experiments was conducted in which the stripping was accomplished by a different method: the sample was cleaved in air and subsequently placed in the deoxygenated solution.

The working solutions were ordinarily 0.1 N KCl, a solution of NaHCO₃ (0.200 g/liter), and also a uranum-containing (approx. $2.5 \cdot 10^{-3}$ g/liter) solution of bicarbonate. These latter solutions were deoxygenated by passing a flow of purified nitrogen through them for 4 h prior to introduction into the cell.

As shown by the data (Fig. 1), the ϕ -potentials in the two series are similar. The rather large scatter in the points for the first series is due to the uneven exposure of the mineral, which depended on the state of the surface of the blade itself, which during the experiments gradually wore down.

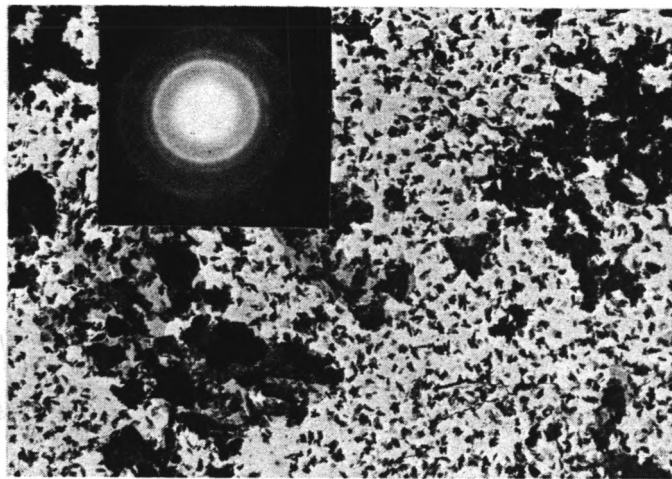
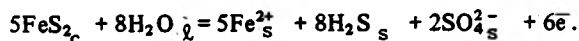


Fig. 4. Very fine particles and films of iron oxide, maghemite $\gamma\text{-Fe}_2\text{O}_3$; formed on the surface of pyrite (sample 8/479) in the experiment with interrupted hermeticity, replica with extraction, magnification 7500. In the inset, their x-ray diffraction pattern.

As a result of the study it was determined that the electrode potential of the freshly exposed surface of all of the pyrite samples studied, which differed in origin, type of conductivity, impurity content (within the range shown) (Table 1), and nitrogen bubbling of the solution, is a value close to -250 mV. The entire set of data points forms a straight line on the summary diagram parallel to the x-axis, i.e., independent of solution pH. This line in acid and near-neutral solutions* can be characterized by the equation for the disproportionation of pyrite in an oxygen-free environment to form ferrous iron, hydrogen sulfide, and sulfate sulfur according to the reaction



The magnitudes of the potentials measured indicate that it is in principle possible for the unaltered surface of pyrites to reduce uranium in an oxygen-free environment.

Below we present experimental results supporting this possibility. For example, in the first series of experiments, a sample of pyrite from the Berezovskii deposit (sample BK) was cleaved in a nitrogen-bubbled bicarbonate solution containing uranium and then held under hermetic conditions for 50 days. At the end of the experiment, the ϕ -potential of the pyrite was -60 ; the pH of the solution was 6.5, and its Eh was $+300$ mV. Electron microscopy showed numerous newly formed orbicular segregations and aggregates of these composed of the uranium oxide UO_2 .

In the second series of experiments, the values measured on a freshly cut surface (sample 1/2203: pyrite concretion from gray clays) after 2 months at the end of the experiments were -155 for the ϕ -potential, 6.5 for the pH, and $+280$ mV for the Eh. In addition to UO_2 , magnetite, wüstite, and newly formed pyrite were detected (Fig. 3).

At the same time, in the second experiment in this series (sample 8/479: pyrite concretion from a gray siltstone) where there was a breakdown in the hermeticity of the system (the experimental conditions were as in the preceding example, except that the ϕ -potential changed from -160 to $+74$ mV in 6 days) only iron oxides were detected on the surface (maghemite, $\gamma\text{-Fe}_2\text{O}_3$, see Fig. 4). And finally, in a control experiment (with water originally containing oxygen), the cut surface of the pyrite (sample 125: concretion from marl) was covered only with oxides and hydroxides of various density and thickness with x-ray diffraction patterns corresponding to wüstite, ferrihydrite, and lepidocrocite (Fig. 5). Thus, there was no

*In the alkaline region, one would expect a pH dependence, which was not observed under the conditions of the experiments, probably because of the screening action of the hydroxide films formed.

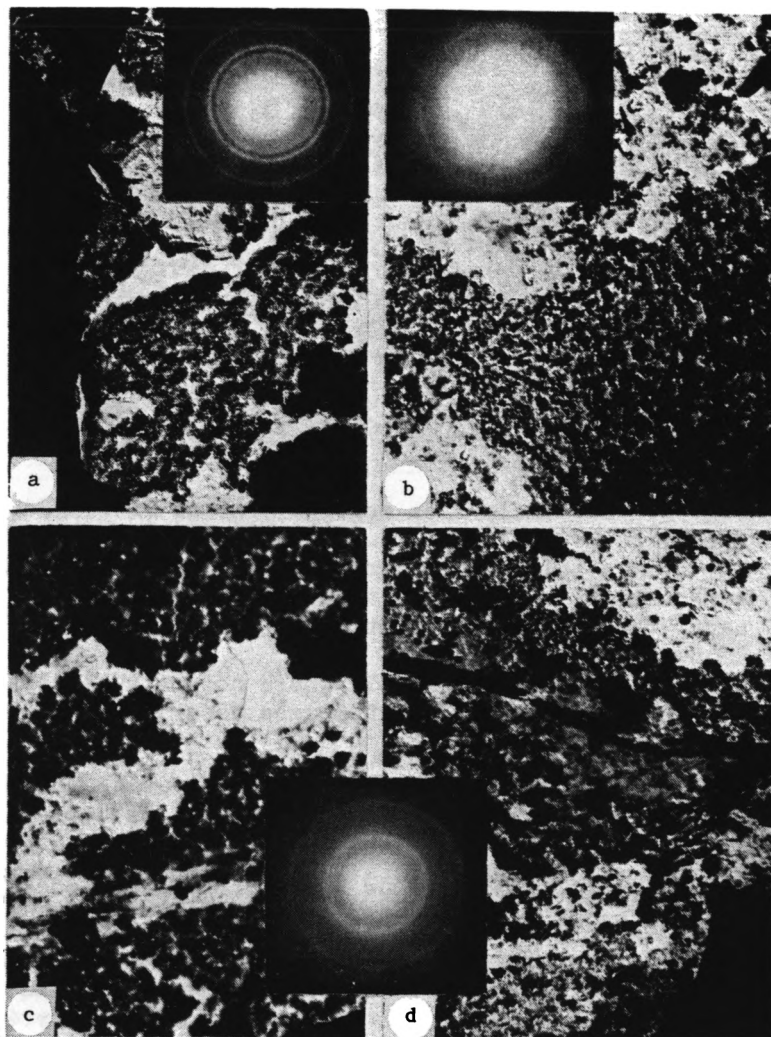


Fig. 5. Films of iron oxides and hydroxides formed in the experiment with water containing oxygen (sample 125), replica with extraction. a) wüstite, magnification 10,500; b) lepidocrocite, magnification 7500; c, d) ferrihydrite, magnification 10,500 and 750, respectively, and x-ray diffraction patterns of these minerals.

reduction deposition of uranium by pyrite under these conditions, probably because if oxygen, a stronger oxidizing agent than U(VI), is also present in the solution, it is preferentially reduced.

The state of equilibrium of the aqueous solution containing pyrite was also determined for the same samples following the method of A. K. Lisitsin [18] in hermetically sealed water-saturated pastes of the powdered mineral (ground in air to 0.1-0.5 mm). This was wet with an oxygen-containing solution of the same composition used in the basic experiments ($\text{UO}_2(\text{NO}_3)_2$ and NaHCO_3 , pH 7.5-7.8).

The measurements showed that the Eh-pH values, as in [8], lay close to the hydrogoethite boundary of the pyrite stability field (Fig. 1). At the end of the experiment, only its oxidation products were found on the surface of the pyrite, hydroxides such as lepidocrocite or ferrihydrite (Fig. 6). No uranium deposition was observed.

The reason for the lack of agreement between the Eh-pH measurements of the water-saturated pastes and the ϕ -potentials is probably mainly that the powdered pyrite (which is more active due to the increase in its surface area and to being ground in air) is strongly screened by oxide films. These insulate the surface of the unaltered pyrite, leading to an increase in potential regardless of the fact that by the time equilibrium is established between the pyrite and the water, the oxygen originally dissolved in the water has been consumed. It is therefore impossible to come to any conclusion about pyrite's ability to reduce

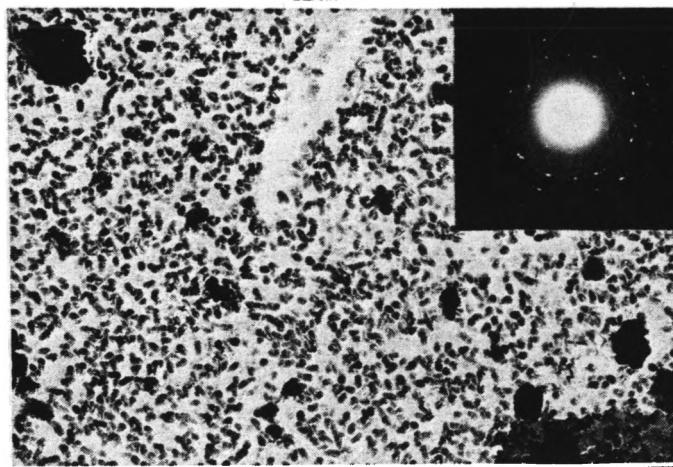


Fig. 6. Extremely fine segregations of ferrihydrite-type iron hydroxide on the surface of pyrite (sample BK) from a cell with water-saturated paste, replica with extraction, magnification 13,500. In the insert, their x-ray diffraction pattern.

uranium based on the potential measured in a cell where no precautions are taken to prevent oxidation of the pyrite. The Eh-pH values in this case describe only the pyrite-goethite equilibrium line.

Thus, these experiments have proved that the unaltered pyrite surface in pyrites of different origins is able to cause uranium to deposit from bicarbonate solutions containing practically no oxygen.

These conditions are observed in nonferrous sulfur-containing rocks beyond the oxidized zones, i.e., in metalliferous areas of oxidizing epigenetic zoning.

These studies have made it possible to resolve a question of fundamental importance for the theory of the formation of exogenic infiltration-type ores, one that has remained a matter of dispute until this time, namely, the possible role of a pyrite reducing geochemical barrier.

LITERATURE CITED

1. A. A. Abramov, "Effect of pH on the state of the pyrite surface," *Tsvetn. Met.*, No. 12, 30-33 (1965).
2. A. A. Abramov, "Electrochemical measurements in studies of mineral surfaces and solutions," *Obogashch. Rud.*, No. 4, 44-45 (1965).
3. E. G. Vasil'eva, "Modeling the processes of the deposition of uranium, selenium, and molybdenum for the interaction of metalliferous oxic waters with extrinsic gaseous reductants," *Litol. Polezn. Iskop.*, No. 6, 54-67 (1972).
4. R. M. Garrels and C. L. Christ, *Solutions, Minerals, and Equilibria*, Harper, New York (1965).
5. S. S. Zavodnov, *Carbonate and Sulfide Equilibria in Mineralized Waters* [in Russian], Gidrometeoizdat, Moscow (1965).
6. I. K. Karpov, A. I. Kiseleva, and F. A. Letnikov, *Computer Modeling of Natural Mineral Formation* [in Russian], Nedra, Moscow (1976).
7. A. V. Kochenov, V. T. Dubinchuk, M. F. Kashirtseva, et al., "Segregation forms and deposition conditions of uranium in exogenic epigenetic deposits," *Geokhimiya*, No. 5, 769-778 (1981).
8. A. K. Lisitsin, *Hydrogeochemistry of Ore Formation* [in Russian], Nedra, Moscow (1975).
9. V. G. Prokhorov, *Pyrite*, Tr. Sib. Nauchno-Issled. Inst. Gidrotechn. Mineral. Syr'ya, No. 103 (1970).
10. R. P. Rafal'skii, "Thermodynamic analysis of equilibria in geochemistry and some conditions for the deposition of uranium in the supergene zone," *Izv. Akad. Nauk SSSR, Ser. Geol.*, No. 4, 96-112 (1978).

DYNAMICS OF THE ACCUMULATION OF SEDIMENTARY
SEQUENCES IN EARLY AND MIDDLE JURASSIC TIME IN
THE NORTHERN CAUCASUS (TRIAL RECONSTRUCTION)

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A. N. Stafeev

UDC 552.14:551.762(470.6)

Data are cited on the rates that sedimentary sequences accumulated during early and middle Jurassic time in the northern part of the Caucasus sedimentation basin. It is shown that sediment accumulation rates were anomalously high, increasing in the following order: shelf-continental slope-slope foot. The effect of such factors as transgressive development of the water basin, eustatic variations of sea level, and others, on the rate of deposit accumulation is examined. We conclude that the significant increase in the total background of sedimentation rates in the eastern part of the basin was connected with the supply there of great masses of sedimentary material by a large river.

One of the most important parameters characterizing the development of sedimentary basins is the rate of sedimentary sequence accumulation. Depending on various factors, it varies within very wide limits, which is attested to by data on contemporary water basins [72]. Consideration of this aspect of ancient sedimentation basins gives interesting results for estimation of both their areal structure and their development in time.

In this article, on the basis of analysis of rates of sediment accumulation, we undertake a trial reconstruction of the dynamics of the development of sedimentary processes in the early and middle Jurassic water basin in the territory of the central and northeastern Caucasus.

One of the main requirements in performing this kind of analysis is a high degree of study of the deposits, their fractional stratigraphic calculation, and comparison over wide areas. The stratigraphic scheme of the lower and middle Jurassic deposits has been worked out by many investigators [2-4, 15, 17-19, 21, and others]; it has recently been significantly refined, and an actually new scheme has been proposed for several regions [5, 6, 9, 16]. This scheme is characterized by the detailed partitioning of sequences. Comparison of various suites in the considered territory is rather reliable, and based on ammonite fauna.

We estimated sediment accumulation rates of the sequences for each of the distinguished suites, and we postulated that the accumulation rate of the deposits was on the whole constant

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sedimentation in that it reflects the character of sediment accumulation over a long period (more than hundreds of thousands of years), during which changes could occur in the sedimentation regime, including erosion of some portion of the already accumulated sediments. Second, ancient deposits were subjected to compaction, particularly in the pure clay varieties. Establishing the degree of rock compaction is an extremely complicated problem, since compaction depends on many factors [28]. Moreover, the magnitudes of RSSA and the sedimentation rates in paleobasins in particular are connected by proportional dependences. Therefore in a series of cases we found it possible to use these concepts as synonyms.

Variations in the structure, composition, completeness of section and other features of sedimentary sequences developed in the Greater Caucasus territory are determined by their correlation to different structural-facies zones (SFZ). In connection with this, we carried out reconstruction of the dynamics of deposit accumulation for different zones. Principles of the structural-facies regionalization are worked out in articles [10, 14-16, 20, and others]. Two large regions are distinguished in the considered region for early Jurassic-Aalenian time: the epigenetic Scythian plate, and the Greater Caucasus geosyncline. They are separated by the Tyrnyauz-Pshekishsk suture zone. A territory is also distinguished in the limits of Balcaria that is limited on the north and south by the Cheget-Dzhorsk and Vladikavkaz faults, and which forms a transition zone (the East-Balkarsk SFZ) between the Scythian plate and the geosyncline (Fig. 1, a).

Several SFZ are established in the limits of the Caucasus mountain-fold structure. In connection with a series of sediment accumulation features in the Central and Northeastern Caucasus, individual zones are distinguished which in space transform one into another respectively: the Digoro-Osetinsk and Gvali-Khivsk SFZ; Pseashkhinsk, Shatili-Dzhurmutsk and the Kudaor-Akhtychaisk SFZ.

Reorganization of the structural plan and corresponding alteration of the structural-facies regionalization occurred in Bajocian-Bathonian time. The following zones are distinguished at this time in the limits of the region of interest (see Fig. 1, b): the Central-Caucasus uplift zone, the Balkaro-Osetino-Dagestansk zone, which is a geosynclinal trough of northern slope, separated by the Sunzhensk dislocation zone from the marginal troughs of the Scythian plate (Baksano-Sulaksk zone).

More than 30 suites are presently distinguished in lower and middle Jurassic deposits in the considered territory, so we therefore did not cite their descriptions and henceforth only utilized the names of the generally accepted and widely known suites.

The authors acknowledge that a series of assumptions were made in this article for conducting reconstruction of the dynamics of sediment accumulation. Henceforth, in proportion to the accumulation of new data (in particular Jurassic reference data), the obtained results may possibly be somewhat altered. However these changes may be concerned with details, and not the conclusions as a whole.

Structure of the RSSA Diagrams. Figure 2 shows the magnitudes of rates of deposit accumulation for sections lying in different SFZ. Calculations of RSSA were conducted for each selected in one or another suite. For convenience in discussion of Fig. 2 in the text, the RSSAs diagram constructed for three actual sections were numbered by roman numerals (I, II, III, etc.), and accordingly the diagrams for each suite are denoted by letters A, B, C, etc., successively from below upwards. In such cases when data was available on changing thickness of some suite from north to south within the limits of a single SFZ, the corresponding RSSA magnitudes were shown under one number on the diagrams and the image acquired a stepped appearance (for example, see Fig. 2, IX, C). We conditionally calculated the RSSA magnitude for suites with insufficiently defined time intervals of their formation or with boundaries not in stratigraphic but in tectonic contact with neighboring suites, and these are shown in Fig. 2 by a "?" mark. For calculation of the diagrams, data on the thickness of sections and their stratigraphic separation were taken from articles [2-7, 9, 14, 18], as well as based on the personal observations of the authors.

RATES OF ACCUMULATION OF DEPOSITS IN DIFFERENT SFZ

The western and eastern parts of the considered region are characterized by somewhat varied conditions of sediment accumulation, and we therefore examine them individually.

In the region of the Central Caucasus, the Jurassic stage of sedimentation began in the limits of the Pseashkhinsk SFZ, most closely arranged at the axial part of the trough. They

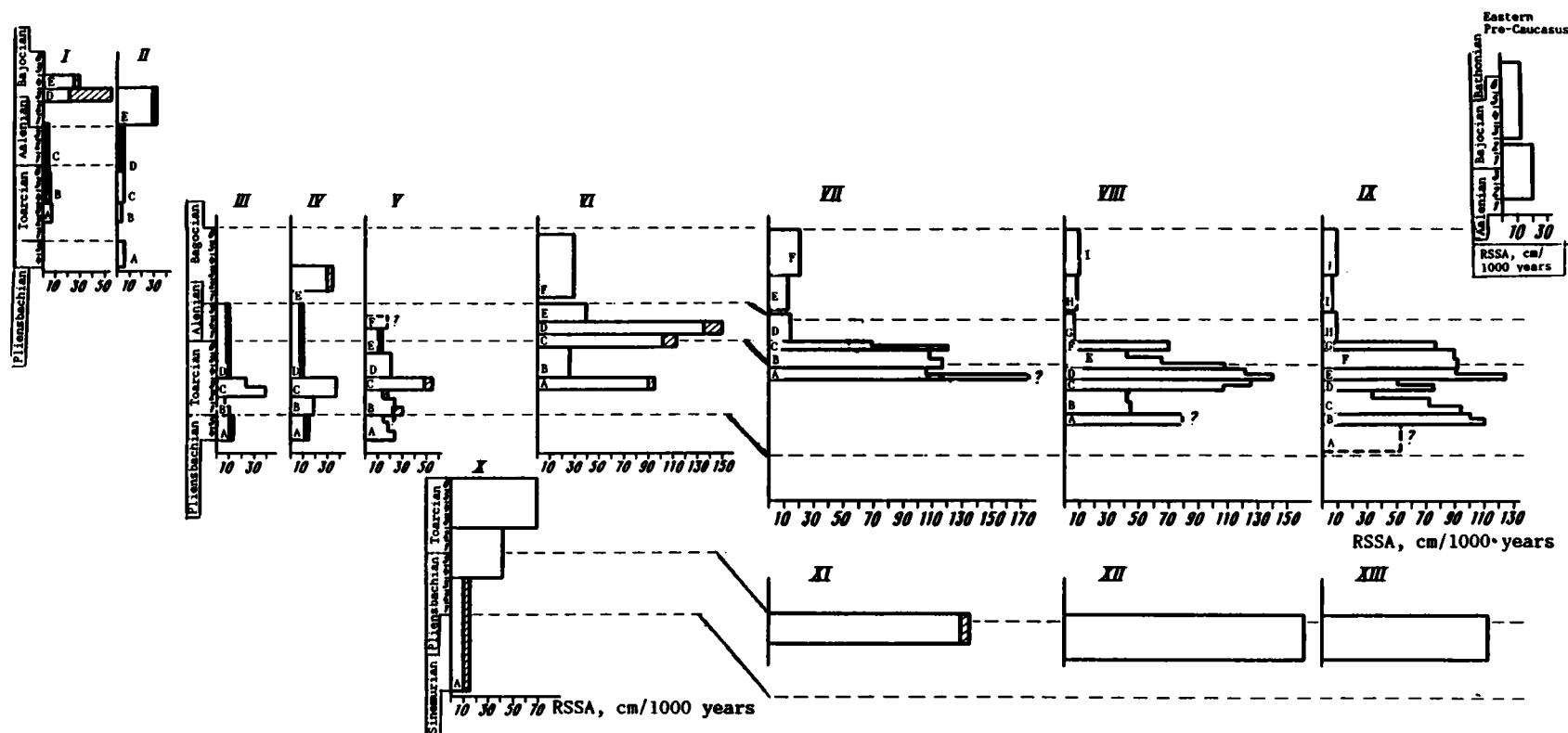


Fig. 2. Diagrams of the rates of sedimentary sequence accumulation. I-XIII) sections: I) Baksan-Tyzyl; II) Cherek Bezingiisk; III) Uruk; IV) Ardon; V) Fiagdon; VI) Assa; VII) Chanty-Argun; VIII) Andiisk Koisu; IX) Avarsk Koisu; X) Terek; XI) headwaters of Chanty-Argun; XII) headwaters of Andiisk Koisu; XIII) headwaters of Avarsk Koisu. Arabic numerals show the zonal division of the stages. Pliensbachian: 1) Uptoni jamesoni; 2) Tragophylloceras ibex; 3) Productylioceras davoei; 4) Amaltheus margaritatus; 5) Pleuroceras spinatum. Toarcian: 1) Dactylioceras tenuicostatum; 2) Harpoceras falciferum; 3) Hildoceras bifrons; 4) Haugia variabilis; 5) Grammoceras thousarsense; 6) Dumorteria levesquei. Aalenian: 1) Leioceras opalinum; 2) Ludvigia murichsonae; 3) Graphoceras concavum. Bajonian: 1) "Sonninia" sowerbyi; 2) Otoites sausei; 3) Stephanoceras humphriesianum; 4) Stenoceras subfurcatum; 5) Garantiana garantiana; 6) Parkinsonia parkinsoni. Slanting hatchures:

were marked by the accumulation of a sequence of coarse deposits: gravelites, coarse and medium grained sandstones, clay-silt shales (Kistinsk suite); its thickness is nearly 1000 m. The granulometric composition and facies aspect of the deposits attest to their accumulation in shallow water conditions, into which the products of the destruction of pre-Jurassic basement were carried from the land. Insufficient determination of the data of the sequence's lower boundary hinders establishment of an accurate time interval for its accumulation. Assuming that formation of the sequence originated over the extent of the entire Sinemurian and early Pliensbachian, its accumulation rates may be estimated at 10-12 cm/1000 years (see Fig. 2, X, A). At the same time, more to the north (in the limits of the Digoro-Osetinsk SFZ) occurred formation of a volcanogenic-sedimentary sequence (the Osetinsk suite), and its accumulation rate may be estimated at no more than 10 cm/1000 years.

Deposits synchronous with the Kistinsk suite in the Northern Caucasus are not exposed in regions lying east of the Terek basin.

With the approach of the Domerian (late Pliensbachian) stage of sediment accumulation, the sea noticeably advanced to north. As a result of changing sedimentation conditions following the deepening of the water basin, in the limits of Pseashkhinsk SFZ the coarse rocks of the Kistinsk suite were replaced by much finer clay-silt deposits with a sharply subordinate quantity of interlayers of fine grained sandstones (the Tsiklaursk suite). The rates of sediment accumulation also grew notably: as seen in Fig. 2, X, B, they are estimated at 40-42 cm/1000 years.

To the north, in the limits of the Digoro-Osetinsk SFZ, which at the time represented the shelf region, that is, a comparatively flat surface sloping somewhat to the east, the clay silts and sands of the Mizursk suite were accumulated. They are comparatively shallow water deposits with numerous layers of creeping animals and bioturbation. Accumulation rates of the sequence reach 12-22 cm/1000 years (see Fig. 2, III-V, A), and were significantly less here in comparison with the same age deposits of the Tsiklaursk suite.

In the Vostochno-Balkarsk SFZ, which was at that time the most northern sedimentation region, the rate of deposit accumulation reached only 5-7 cm/1000 years (see Fig. 2, II, A), that is, roughly 8 times less than in the Pseashkhinsk SFZ. Thus, comparison of the RSSA diagrams for the Domerian stage shows a distinct tendency toward increase in the sediment accumulation rates from the shallow water near coastal regions to the deeper water part of the water basin lying close to the axis. Considering that in compositional distinction from the rocks of the Mizursk suite, the rocks of the Tsiklaursk suite are much more argillaceous and were correspondingly exposed to post-sedimentational compaction to a greater degree, the differences in sedimentation rates were still greater than in comparison of the thicknesses of the compacted rocks.

In the early Toarcian (ammonite zones tenuicostatum and falciferum), formation of the Tsiklaursk suite proceeded in the Pseashkhinsk SFZ (40-42 cm/1000 years). This was the most argillaceous part. The sea temporarily receded from the territory of the Vostochno-Balkarsk SFZ in this period (there are no deposits here for this time). Sediment accumulation was prolonged in the Digoro-Osetinsk SFZ and predominantly argillaceous deposits were formed. Moreover we note that textural-structural signs of the rocks indicate the existence here of completely unstable sedimentation conditions in a shallow water basin. As seen from Fig. 2, III-IV, B, the RSSA magnitude in various sections varied from 6 to 32 cm/1000 years, increasing in an easterly direction. However this increase is apparently connected with the fact that the more eastern sections in this SFZ characterize sediment accumulation at great distance from the coast, where the rates were higher.

At the end of early Toarcian and throughout late Toarcian time in the Pseashkhinsk SFZ formation of a predominantly clay-silt sequence of marine suite occurred. In comparison with the preceding stage, the RSSA magnitude increased by almost 30 cm/1000 years, reaching 70 cm/1000 years (see Fig. 2, X, C). It is evident that the RSSA magnitude varied during accumulation of sequences, but insufficient faunal characterization of the deposit did not make it possible to distinguish more fractional stratigraphic subdivisions.

In every SFZ lying north of the Pseashkhinsk SFZ, the end of the early Toarcian (ammonite zone Hildoceras bifrons) was notably distinct from other stages of sedimentation. Northward transgression of the sea occurred at this time*: sediment accumulation in the Vostochno-

*From V. P. Kazakova's data [8], the start of the transgression occurred in the zone Harporceras falciferum.

Balkarsk SFZ was renewed, accumulation of marine deposits in the limits of the Labino-Malkinsk SFZ began, and the sea first covered the southern part of the Scythian platform and formed a geomorphologically distinctly expressed shelf region.

Analysis of RSSA diagrams show that relatively increased rates of sediment accumulation correspond to this time everywhere, although they also varied significantly in different SFZ. If the RSSA magnitudes in the Labino-Malkinsk and Vostochno-Balkarsk SFZ are respectively equal to 7 and 5 cm/1000 years (see Fig. 2, I, A, II, B), then in the Digoro-Osetinsk zone they increase to 40-55 cm/1000 years (see Fig. 2, III-V, C), and in the Kestantinsk basin (Pshekish-Tyrnyausk suture zone) they increase up to 50 cm/1000 years.

In late Toarcian and Aalenian time, a slowing of accumulation rates is generally noted (within the limits of the shelf and on the continental slope) in comparison with the preceding stage. In the limits of the Digoro-Osetinsk SFZ, the RSSA magnitudes for different sections and for different time intervals vary from 8 to 21 cm/1000 years (see Fig. 2, III-IV, D, V, D-E), in the Labino-Malkinsk and Vostochno-Balkarsk zones they reach 2-6 cm/1000 years (see Fig. 2, I, C, II, D); in the Kestantinsk depression they reach 12-30 cm/1000 years.

The following Bajocian stage of development in the Jurassic water basin is occupied by a notably increased RSSA magnitude, which reached 32 cm/1000 years in the territory of the former Digoro-Osetinsk SFZ, while in the Labino-Malkinsk and Vostochno-Balkarsk SFZ the rates are 30-58 cm/1000 years. This is connected with the tectonic reorganization of the Greater Caucasus region and significant alteration of the basin morphology. This caused shifts in the regions of maximal sediment accumulation rates in the water basin area.

The RSSA diagrams of the eastern part of the considered Jurassic water basin region are distinguished from the western part by a series of features. First of all the overall increase in the RSSA magnitude for Lias-Aalenian time draws attention. Thus, for the Shatili-Dzhurmutsk SFZ, RSSA values increase in comparison with same aged formations of the Pseashkinsk SFZ from 40 to 160 cm/1000 years (see Fig. 2, XI-XIII). For sections of the Agvali-Khivsk SFZ, the configuration of the RSSA diagram is such that the overall high background rates of sediment accumulation may resolve into two relative maxima (see Fig. 2, VI, A; VIII, A; IX, B). The lower, reaching 110 cm/1000 years, corresponds to the end of the early Toarcian (ammonite zone H. bifrons) and is associated with a similar maximum in the western part of the profile. Thus, it has at least a regional character. The other extreme RSSA (up to 120-140 cm/1000 years) is observed in the upper Toarcian and corresponds to the lower part of the ammonite zone Dumorteria pseudoradiosa (Assabsk suite [5]); it is clearly distinguished in sections of Western Dagestan and Chechna.

The significant decrease in the RSSA (6-17 cm/1000 years) in the late Aalenian-start of early Bajocian (see Fig. 2, VII; D; VIII, G; IX, H) draws attention, and high RSSA values (40 cm/1000 years) were characteristic only in the section of the Assa river for this stage (see Fig. 2, VI, E).

In the limits of the wide Agvali-Khivsk SFZ in several cases it is possible to observe sections of single aged deposits lying transverse to the strike of the zone, that is, at different distances from the shore line of the paleo water basin. It is noted that the sections of such suites that are located at great distance from the shore line (more to the south) invariably seem thicker than in their northern analogs. This circumstance is directly connected with the increased rates of sediment accumulation in the southern part of zone, and found reflection in the RSSA diagrams: as seen from Fig. 2, IX, the difference between them may reach 20-60 cm/1000 years.

In late Aalenian time, as a result of transgressions, the sea advanced far to the north into the territory of the Eastern Pre-Caucasus, which formed the eastern part of the water basin's shelf. Sediment accumulation for the shelf region took place here at rather high rates: the RSSA magnitude for the late Aalenian-start of early Bajocian locally reached 20 cm/1000 years, that is, notably higher in comparison to the western shelf (Labino-Malkinsk SFZ).

Low rates of sediment accumulation were characteristic for the eastern part of the region in Bajocian time in comparison with Lias. Interesting features inherent to this stage of sedimentation were detected: the RSSA magnitude increased from east (6-12 cm/1000 years, see Fig. 2, VIII, I; IX, J) to west (13-27 cm/1000 years, see Fig. 2, VI, F; IV, E). They nonetheless remain less than in the Digoro-Osetinsk SFZ. Thus, a reverse picture of the distribution of RSSA magnitudes is observed in Bajocian time (in comparison with Lias).

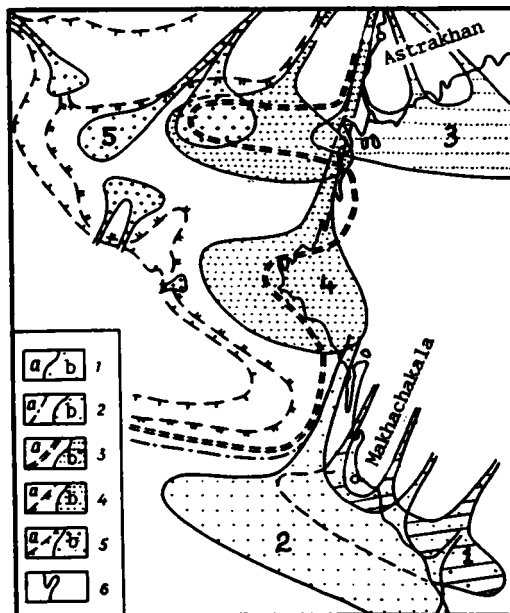


Fig. 3. Location of the coast line (a) and paleoriver delta (b) in the northeastern part of the Greater Caucasus basin. 1) middle of late Toarcian (proposed position); 2) end of late Toarcian-start of the late Aalenian; 3) late Aalenian-start of early Bajocian; 4) early Bajocian; 5) late Bajocian-start of Bathonian (?); 6) shore line of the Caspian sea.

In the limits of the Eastern Pre-Caucasus rates of sediment accumulation were also not great. The RSSA value did not exceed 5-12 cm/1000 years, and rates were much less than in the Labino-Malkinsk SFZ (30-50 cm/1000 years).

FACTORS INFLUENCING RATES OF SEDIMENT ACCUMULATION IN THE EARLY AND MIDDLE JURASSIC WATER BASIN

Analysis of available data indicate that rates of sediment accumulation in the early and middle Jurassic geosynclines of the Greater Caucasus were extremely high. As shown earlier, RSSA magnitudes in the shelf zone varied within limits from 4 to 20 cm/1000 years, on the continental slope they increased, and reach 70-90 cm/1000 years, and even more than 120 cm/1000 years, and in the zone corresponding to the slope foot, the accumulation rates are maximal: 120-160 cm/1000 years. Considering that we operate on data from compacted rocks, the actual rates of sediment accumulation were 2-3 times higher.

A. P. Lisitsyn [13], analyzing contemporary environments of sediment accumulation, arrived at the conclusion that it is necessary to consider rates of sedimentation exceeding 10 cm/1000 years as anomalously high (avalanche sedimentation). In agreement with this criterion, in the Jurassic water basin the rates of deposit accumulation on the continental slope and at its foot, as well as locally in shelf regions, are anomalously high, sometimes exceeding it by more than 10 times.

The overall high background of rates of sequence accumulation is complicated by the extreme unevenness in the distribution of RSSA values. The origin of this unevenness is caused by both the overall relationships in the development of the entire geosyncline of Greater Caucasus, and by local causes. In addition it is necessary to note that the Jurassic water basin existing in the territory of the Greater Caucasus was part of the Tethyan ocean, and accordingly reflected the defined tendencies in its development.

One of the basic factors influencing to some degree the character of sediment accumulation in the geosyncline of the Greater Caucasus was the transgressive sea which was developing over the span of the early and middle Jurassic [3, 15, and others]. This was closely connected with downward movement in the trough zone. In fact, if we compare the distribution of deposits of various age, then we see that beginning with the Sinemurian, younger deposits occupied a significantly wider area than formerly. Thus, for example, in the considered territory, if deposits of the Kistinsk suite in Sinemurian time occurred within the limits

of the Pseashkhinsk SFZ, then distribution region of the upper Aalenian sequence included the southern part of the Scythian plate, and in the Bajocian they advanced still more to the north.

It is necessary to note that in proportion to the overall transgressive development of the early and middle Jurassic basin over the course of time the "geomorphological function" of one or another portions of the marine floor were altered, and this entailed alteration in the character of the sediments and their accumulation rates. Thus, the siltstone and sandstone deposits which were accumulated in Domerian time in the limits of the Digoro-Osetinsk SFZ were characteristic of shelf deposits in their facies aspect. The fact of the uniform character of deposits proceeding over several tens of kilometers transverse to the strike of the zone indicates the rather large width of the paleoshelf. Later, for example at the end of the early Toarcian, this territory was intensively deflected and transformed into a continental slope with higher rates of sediment accumulation than before, while the southern part of the Scythian plate began to fill the function of the shelf at this time. Accordingly, significant variations in RSSA magnitudes are observed in the actual sections (see Fig. 2) for these two intervals.

The uneven development of the transgression had an effect on the character of sediment accumulation: periods of intensive advance of the sea onto dry land alternated with epochs of its relative stabilization or even absence. Instability in the position of the coast line affected activity in the delta of the large river lying in the east of the region [22]. Introduction of a significant mass of sedimentary material by this river caused an increase in the RSSA magnitude here, which is the cause of the clear asymmetry of the profile (see Fig. 2). Since the asymmetry began to be apparent by at least Domerian time,* it is clearly possible to consider that the delta already existed at this time. Territorially it was evidently located in the region of the western coast of the contemporary Caspian (Fig. 3).

It is interesting that in late Toarcian-early Aalenian time in the limits of the Eastern Caucasus the deltaic facies advanced far to sea (see Fig. 3), forming thick principally sandstone sequences with thin mud layers (Karakhsk [22], Batlukhst, and Datunsk suites [5]). V. T. Frolov connected the advance of the delta in the direction of the sea with the rising movements of dry land, which were significant in the territory of the Northern Caspian. Moreover lithological investigations of lower and middle Jurassic deposits in the limits of the Central and Western Caucasus [1] indicate shallowing of the sea at the end of the Toarcian era. This was also earlier connected with the local revival of rising movements at this time. Considering the records of the fall in the level of the water basin which is noted over wide areas, it is possible to consider with sufficient confidence that a regressive stage in the development of the Jurassic water basin in the Caucasus is connected with this time. Paleographic analysis of the Jurassic period conducted by E. Khellem showed that inherently eustatic lowering of sea levels was general at the end of late Toarcian-beginning of Aalenian time [25, 26]. In connection with this, we represent the regression in the Northern Caucasus as the reflection of a wider event.

Lowering of the level of the Caucasus water basin led to the fact that earlier accumulated delta deposits were beginning to be actively eroded and redeposited at a new level which was forming in a region where a marine deposit sequence of deltaic facies was developing. It would have been possible to anticipate appearance of particularly high rates of accumulation of deltaic deposits in the section in comparison with the remnant purely marine portion of the sequence. However, as seen from Fig. 2, IX, F, G, although the RSSA values are also great, they do not exceed, for example, the preceeding. This situation is connected with the fact that deposits accumulating in the region of the developing delta and fore-delta, as a further consequence of active hydrodynamics, were unevenly eroded and repeatedly redeposited, which is indicated by multiple traces of underwater erosion of the sediments. Moreover, the sequence underlying the deltaic complex and representing deposits of the underwater output cone, which were accumulated in a stable environment, are characterized by the highest RSSA values. It is entirely probable that synchronous deltaic accumulation sequences (see Fig. 2, VI, C, D) also characterize deposits of the underwater output cone with the high RSSA magnitudes inherent to it.

In the Central Caucasus where such a large supplier of sedimentary material did not exist, as it did in the east, and accordingly there were no large deltaic accumulations,

*More ancient deposits on the east of the Greater Caucasus are not detected in the limits of the considered region.

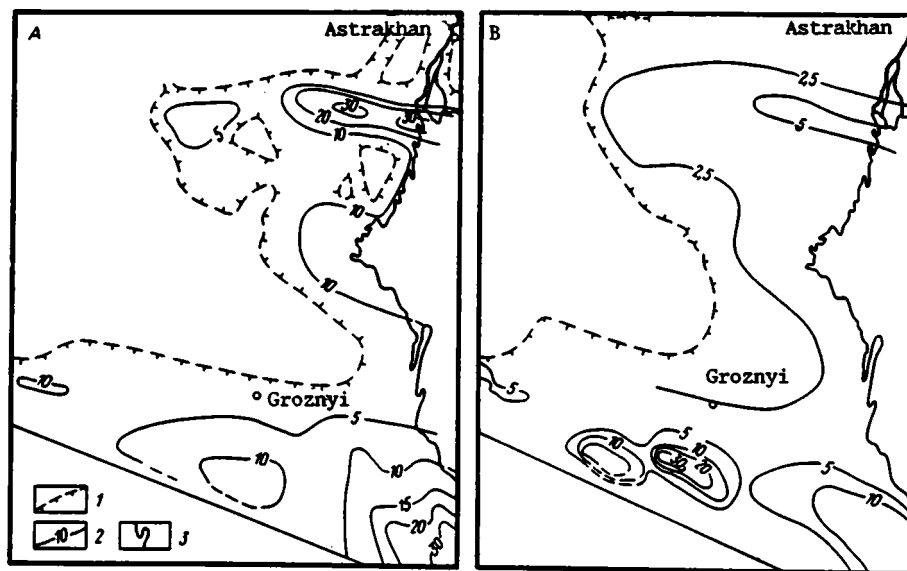


Fig. 4. Scheme of the rates of deposit accumulation for the late Aalenian-start of early Bajocian (A) and late Bajocian-start of early Bathonian (B) time. 1) position of the coast line; 2) isolines of rates of deposit accumulation; 3) position of the coast line of the Caspian sea.

the late Toarcian-early Aalenian regression did not give rise to a notable redistribution of the sediments. Conversely, shallowing of the sea and strengthening of the hydrodynamics did not assist in the accumulation of sedimentary material in the shelf and continental slope zones. This was carried away to the deeper water part of the basin and this apparently also caused the decrease in the RSSA values in these regions.

In addition to the late Toarcian-early Aalenian regression, traces of intermittent absence of the sea are also noted in some stratigraphic levels of the J_{1-2} deposits of the Northern Caucasus; thus, for example part of the Lower Toarcian sequence in the Vostochno-Balkarsk SFZ and Arkhyz-Guzeripl'sk SFZ of the Western Caucasus [11] is absent from the section. These rising movements are considered a local effect. However the effect in single-aged deposits of Dagestan (Ratlubsk suite [5]) of lenses and interlayers of coarse material in Dagestan (in particular, conglomerate-like lenticular interlayers), may apparently indicate the much wider influence of rising movements, and corresponding regression and erosion of more ancient rocks accompanying the movements. Unfortunately, incompleteness in the section of the Ratlubsk suite hinders a more or less accurate estimate of the RSSA magnitude for Dagestan (see Fig. 2, IX, A); in the limits of the Central Caucasus the rate of deposit accumulation at this time (and in the late Toarcian-early Aalenian) was comparatively less (see Fig. 2, III-V, A).

These regressive episodes were varied by clearly expressed transgressive stages, when the sea notably advanced to north. These transgressions in the second half of the Toarcian and in the late Aalenian significantly influenced sediment accumulation in the geosynclinal basin. The Aalenian transgression had a particularly significant aftereffect: as seen in Fig. 2, IX, rates of sediment accumulation at this time were sharply decreased (by 7-10 times in comparison with preceding period). What caused such a significant slowing in the accumulation rates of the sedimentary material?

As a result of the swiftly proceeding transgression the vast territories of the Eastern Pre-Caucasus were submerged and the shore line here advanced to the north by roughly 300 km. Migration of the paleo-Volga delta occurred correspondingly. Its position in late Aalenian time varied significantly in comparison with the early Aalenian (see Fig. 3). As a result of the transgression, a wide shallow water shelf formed, where numerous rewashed clay-silt and sand deposits accumulated. On the whole, sediment accumulation in the territory of the Eastern Pre-Caucasus occurred at comparatively low rates (RSSA 10 cm/1000 years) (Fig. 4, A). Only in the northern part of this region did the suture zone of the East European platform and the Scythian plate undergo notable subsidence, and a significant quantity of terrigenous

material was collected and accumulated with the highest rates (RSSA 30 cm/1000 years) in comparison with other parts of the shelf in the forming gulf. The material was brought by the paleo river system from the East European platform. Some part of the sedimentary material was carried away in the water basin from a system of mid-mountain uplifts existing on the territory of the Scythian plate. Their height was no less than 0.6-1 km, based on palynological data.

In spite of the relatively low rates of sediment accumulation, a significant mass of sedimentary material was collected and buried here over a huge area. Accordingly, only a relatively small quantity of sediments taken by rivers from the north to the water basin reached the geosynclinal trough. This circumstance was one of the main causes of the significant slowing in the rates of sediment accumulation within the limits of the Agvali-Khivsk SFZ. The RSSA magnitudes here over a significant territory did not exceed 10 cm/1000 years, and they only increased in a southeast direction (see Fig. 4), and in a western direction where Pre-Caucasus land began to play the role of sedimentary material source. But on the whole it is possible to consider that in late Aalenian-early Bajocian time rates of deposit accumulation in the Eastern Pre-Caucasus territory and on the northern flank of the Eastern Caucasus trough were approximately equal.

Thus, slowing of the rates of sedimentary sequence accumulation in the region of the continental slope was basically caused by displacement of the trough (proceeding as a result of transgression) to a significant distance from the delta which was the main supplier of sedimentary material in this part of the region.

Alternatively, transgressions influenced sedimentation in the second half of early Toarcian time. Significant increases in the rates of deposit accumulation coincided with them. This interval was notably distinct among others in terms of RSSA values, particularly within the limits of the Central Caucasus (see Fig. 2, III-V, C), and to a lesser degree to the east because of the generally high background RSSA magnitudes here.

The sea advanced to the north by roughly 25-30 km within the limits of the Labino-Malkinsk SFZ in the Central Caucasus; it is more difficult to carry out a direct estimate in the eastern regions, where the sea's advance was less due to the high shoreline. The transgressions led to shifting of the sedimentation basin's depocenter to the north, or at least to the significant expansion of the zone of high RSSA values and their distribution within the limits of the Digoro-Osetinsk and Agvali-Khivsk SFZ, while they were earlier correlated to regions attracted to the axial part of the water basin. As a result of the transgression and formation of a marine shelf in the southern part of the Labino-Malkinsk SFZ, the Digoro-Osetinsk SFZ was itself transformed into a slope and slope foot region, and became a zone of active sediment accumulation. Further, in proportion to the accumulation of a sedimentary sequence here and the prolonging the subsidence of the axial part of the water basin, the maximum of sediment accumulation rates again became displaced to the south and the RSSA magnitude notably declined in the Digoro-Osetinsk SFZ.

Thus, considering sediment accumulation in similar SFZ during two stages of the sea's advance onto dry land, it is necessary to note that first, the transgressions undoubtedly had a significant effect on the sedimentation process; second, these phenomena mutually existed with one another in complex interconnection and there was no clearly expressed proportional dependence between them; if slowing of sediment accumulation rates occurred in one case, then there was conversely an increase in the other.

Structural reorganization of the region took place at the start of early Bajocian time [15, and others]. In some areas this process was accompanied by the sea's withdrawal and partial erosion of earlier accumulated sequences. Differentiation of the basin into a series of new structural-facies zones occurred as a result of this reorganization (see Fig. 1). A system of central Caucasian uplifts was formed in the axial part of the geosynclinal trough and these bounded the zone of the Balkaro-Osetino-Dagestansk trough from the south. Reorganization of the region's structural plan caused displacement of the basin's depocenter of sedimentation to the north; it now lay roughly on the territory of the northern part of the former Digoro-Osetinsk and Agvali-Khivsk SFZ, and possibly in the zone of their suture with the Scythian plate (see Fig. 4, B).

Uneven rates of sediment accumulation in various portions of the water basin were characteristic for Bajocian-Bathynian time (as well as for the late Aalenian). In the new stage in the limits of the vast territory of the Eastern Pre-Caucasus the rates of sediment

accumulation slowed somewhat in connection with the further migration of the paleo-Volga delta to the north. However, as previously, some great rates of sediment accumulation were distinguished, although weakly, in the suture zone of the East-European platform and the Scythian plate (see Fig. 4, B).

There is a clear asymmetry in the distribution of RSSA magnitudes in the limits of the zone of the Balkaro-Oestino-Dagestansk troughs. But in distinction from Lias-early Aalenian time, when a clear tendency was reorganized towards an increase in the rates of sediment accumulation in an eastern direction, the opposite picture was observed in Bajocian time: greater RSSA magnitudes are correlated to the western part of this zone (see Fig. 2). Such a distribution of RSSA values is similar to late Aalenian time and has the same cause: a sharp decrease in the quantity of sedimentary material supplied to the eastern part of the trough. At the same time, to the west the water basin was significantly constructed and sedimentary material could enter from both the north (from the region of Pre-Caucasian uplifts) and from the south (from the zone of central uplifts). This caused the rather high rates of sediment accumulation here. The existing system of faults cutting the Caucasian structure led to the division of the basin into a series of depressions, which were restricted in their RSSA magnitudes.

Summing up the discussion, we note the following:

On the whole, the sedimentary process in the Caucasus basin was characterized by anomalously high sedimentation rates. Rates of sedimentary sequence accumulation in different parts of the early and middle Jurassic water basin were determined to a significant degree by the morphostructural elements of the basin of sediment accumulation which were correlated to one another: the smallest RSSA magnitudes were intrinsic to the shelf regions, they increased in the direction: continental slope → slope foot → northern setting of the axial part of the trough. The transgressive developmental trend of the Jurassic basin accompanying its gradual filling by sedimentary material led to the fact that a definite part of the floor of the water basin occupied different geomorphological functions in different time periods (for example, the region of the Domerian shelf was transformed into continental slope in Toarcian time). Accordingly, rates of sediment accumulation in these regions varied in the course of time. Variations in rates of sediment accumulation in different parts of the trough, which led to the asymmetrical distribution of RSSA magnitudes in its central and eastern parts, were caused to a significant degree by the activity of large rivers bringing a great quantity of sedimentary material to the water basin. In periods of slowing of the downwards motions, the accumulating sedimentary sequences occupied several parts of the water basin, significantly changing the morphology of its floor, affecting the sea's depth, and accordingly the subsequent environments of sediment accumulation. That is, a strictly sedimentary process determined and regulated the conditions of sedimentation (including the rate) to a significant degree in various stages of the basin's development.

LITERATURE CITED

1. R. S. Bezborodov, "Lithology of upper Lias and middle Jurassic deposits of the central part of the northern slope of the Caucasus in connection with prospects of their oil and gas potential," in: *Geology and Oil and Gas Potential of the Southern USSR* [in Russian], Gostoptekhizdat, Leningrad (1961), pp. 107-254.
2. N. V. Beznosov, *Bajocian and Bathonian Deposits of the Northern Caucasus* [in Russian], Nedra, Moscow (1967).
3. N. V. Beznosov, V. P. Kazakova, Yu. G. Leonov, and D. I. Panov, "Stratigraphy of lower and middle Jurassic deposits of the central part of the Northern Caucasus," in: *Material on the Geology of Gas Bearing Regions of the USSR* [in Russian], Gostoptekhizdat, Moscow (1960).
4. N. V. Beznosov and V. V. Shelkhovskii, "The Jurassic system. Lower and middle divisions. Eastern part of the Northern Caucasus," in: *Geology of the USSR* [in Russian], Vol. IX, Part I, Nedra, Moscow (1968), pp. 168-185.
5. A. I. Gushchin, "Conformity of geological development of the Northeastern Caucasus in early and middle Jurassic time," Candidate's Dissertation, Geologic-Mineral Science, Moscow State Univ. (1986).
6. A. I. Gushchin and D. I. Panov, "On the stratigraphy of lower Jurassic deposits of the anticlinorium of the Bokov range (Dagestan)," *Vestn. Mosk. Gos. Univ., Ser. 4, Geol.*, No. 5, 19-28 (1983).
7. V. P. Kazakova, "Towards a stratigraphy of lower and middle Jurassic deposits of the Aigamuga-Don river basin (Gornaya Osetiya)," *Izv. Vyssh. Uchebn. Zaved., Geol. Razvedka*, No. 8, 60-65 (1958).

8. V. P. Kazakova, "Toarcian hildoceratidia (ammonoids) from the Dzhigiatsk suite of the interriver Bol'shoi Zelenchud-Kuban' (Northern Caucasus)," Byull. MOIP. Otd. Geol., 62, No. 1, 86-102 (1987).
9. V. P. Kazakova, A. I. Gushchin, and D. I. Panov, "Upper Pliensbachian ammonites and the age of lower horizons of the lower Jurassic deposits in the Eastern Caucasus (in Dagestan territory)," Byull. MOIP. Otd. Geol., 61, No. 4, 61-78 (1986).
10. Yu. G. Leonov, "Structural-facies zonation of the early Jurassic-Aalenian trough of the Central and Western Caucasus," Dokl. Akad. Nauk SSSR, 167, No. 1, 166-169 (1966).
11. Yu. G. Leonov, "Early and middle Jurassic phases of uplift and fold formation in the Greater Caucasus," Geotektonika, No. 6, 31-38 (1969).
12. A. P. Lisitsyn, Sediment Formation in the Oceans [in Russian], Nauka, Moscow (1974).
13. A. P. Lisitsyn, "Avalanche sedimentation in seas and oceans. Report 1," Litol. Polezn. Iskop., No. 6, 3-27 (1983).
14. D. I. Panov, "Stratigraphy, magmatism, and tectonics of the Greater Caucasus in the early Alpine stage of its development," in: Geology of the Greater Caucasus [in Russian], Nedra, Moscow (1976).
15. D. I. Panov, "Structural-facies regionalization of the Greater Caucasus in the early Alpine stage of its development (early and middle Jurassic)," Byull. MOIP. Otd. Geol., 63, No. 1 13-24 (1988).
16. D. I. Panov and A. I. Gushchin, "Structural-facies regionalization of the territory of the Greater Caucasus for early-middle Jurassic time and the regional-stratigraphic partition of lower-middle Jurassic deposits," in: Geology and Mineral Resources of the Greater Caucasus [in Russian], Nauka, Moscow (1987), pp. 124-139.
17. V. M. Pats, "Towards a Jurassic stratigraphy on the Chanty-Arguni river (Checheno-Ingush ASSR)," Tr. po Geol. i Polezn. Iskop. Severnogo Kavkaza, No. 1, 153-161 (1938).
18. Yu. Ya. Potapenko and V. N. Beznosov, "The Jurassic system. Lower and middle divisions. Central part of the Northern Caucasus," in: Geology of the USSR [in Russian], Vol. IX, Part I, Nedra, Moscow (1968).
19. V. P. Rengarten, "Geologic outline of the Voennno-Gruzinsk route," Tr. Vsesoyuz. Geol.-Razved. Ob'edineniya, No. 138 (1932).
20. K. O. Rostovtsev and L. A. Nikanorova, "The stratigraphy and basic outline of the tectonic development of the Greater Caucasus and Pre-Caucasus in the early and middle Jurassic," Sov. Geol., No. 5, 3-19 (1970).
21. I. D. Filimonov, "Short geologic outline of the Andiisk Koisu river basin in Dagestan," Tr. po Geol. Polezn. Iskopaemym Sev. Kavkaza, No. 1, 129-152 (1938).
22. V. T. Frolov, Trial and Methods of Complex Stratigraphic-Lithologic and Paleogeographic Investigations [in Russian], Moscow State Univ. (1965).
23. U. B. Kharlend, A. V. Cox, P. G. Llewellyn, et al., The Scale of Geologic Time [Russian translation], Mir, Moscow (1985).
24. A. Hallam, The Jurassic Period [in Russian], Nedra, Leningrad (1978).
25. A. Hallam, "Eustatic cycles in the Jurassic," Paleogeogr., Paleoclimatol., Paleoecol., 23, No. 1, 1-32 (1978).
26. A. Hallam, "A revised sea-level curve for the early Jurassic," J. Geol. Soc., 138, Pt. 6, 735-744 (1981).
27. D. V. Kent and F. M. Gradstein, "A Cretaceous and Jurassic geochronology," Bull. Geol. Soc. Am., 96, No. 11, 1419-1427 (1985).
28. H. H. Rieke and G. V. Chilingarian, Compaction of Argillaceous Sediments, Elsevier, Amsterdam (1974).

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Lithologic, facies, mineralogic, and petrographic features of carbonate evaporitic sediments of Syria are described. For the first time a genetic classification of these rocks is proposed. It is used to reconstruct the paleogeography of the platform during the Triassic.

The Triassic gas and oil rock complex which is productive in the area of the Palmyrides, the eastern slope of the Aleppo arch, the Euphrates depression, and northeastern Syria (Rumelan and Jebesa districts), is quite widespread in the territory of Syria.

All proven hydrocarbon deposits are confined to carbonate sediments of the Kurachina-dolomite suite. The deposits are screened by evaporitic rocks (anhydrites and rock salt) of the Kurachina-anhydrite and Aday suites.

Studies of the Syrian Triassic suggest a complex structure of these rocks due to facies variability and development of zones and subzones making up the pool. The productive Kurachina-dolomite suite is mostly represented by dense dolomites with sporadic porous and fissured varieties. The variation of fissured collectors suggests pronounced homogeneity of main collector properties. In several deposits, the best collectors (with porosity up to 5-7%) occur mainly in near-arch parts of structures; in wings their properties greatly deteriorate.

Until recently all proven oil and gas accumulations in the Syrian Triassic were associated with anticlinal traps, with diversity of reservoirs, folded structures, hydrocarbon compositions, well productivity, etc. Nevertheless, future prospects of oil and gas in the Triassic should not be limited to structural traps, which are being rapidly exhausted. Data indicate the possibility of nonanticlinal deposits in this region, associated with lithologic traps due to local development of fissured collector zones, stratigraphic discontinuities, and fault ruptures, as well as combination type traps.

To this end, detailed lithologic-facies and mineralogic-petrographic investigations of the Triassic in Syria conducted in 1983-85 and 1987 by joint efforts of Soviet and Syrian geologists have established the genesis and formation conditions of the above rock complex with a view to possible prospecting for nonanticlinal traps. The study was based on lithologic-facies analysis developed and updated at the Geological Institute of the USSR Academy of Sciences [5, 11].

The Triassic is exposed on limited areas in the western part of the country. The study mostly relied on deep well drilling (Fig. 1). Lithologically, the Triassic rocks are highly diverse. They are subdivided into 10 suites (Fig. 2). Genetically, they appear as a complex-structured multifacies strata comprised of a succession of continental littoral-marine and marine complexes.

Lithologic and Petrographic Characteristics of the Triassic

The Triassic consists of carbonate, sulfate, and terrigenous entities. Lithologic and facies studies of these bodies have shed light on their genesis and the history of the Triassic sedimentation basin.

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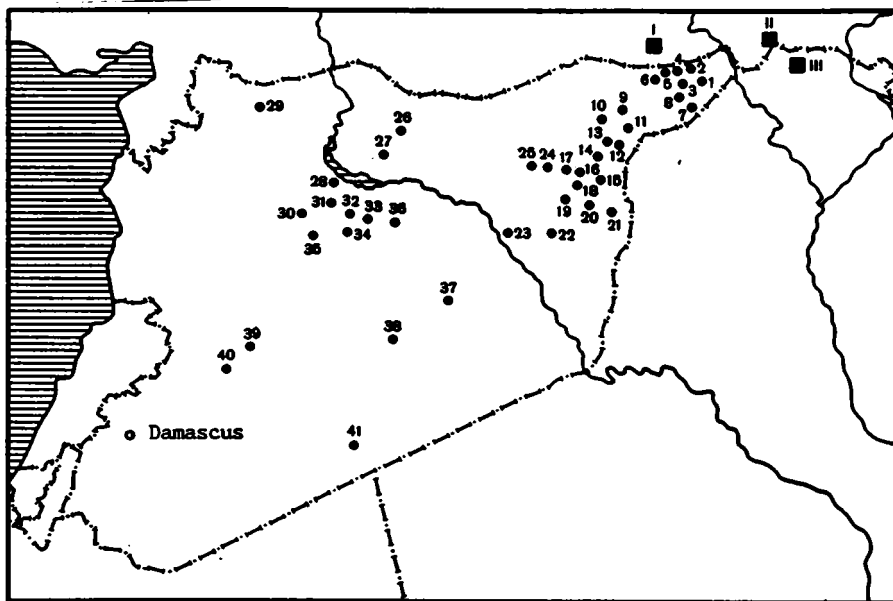


Fig. 1. Distribution of deep wells and sections. Wells: 1) Rumelan-6, 2) Alayan-101, 3) Hadadi-1, 4) Badran-1, 5) Aoda-107, 6) Shamo-1, 7) Naur-1, 8) Hrbet-3, 9) Badri-101, 10) Sheik-Mansur-1, 11) Nora-1, 12) Khol-1, 13) Abu-Haydar-1, 14) Tishrin-5, 15) Gbebi-101, 16) Jibisi-208, 17) Salhiya-1, 18) Rim-1, 19) Mazi-1, 20) Markada-101, 21) Bahri-1, 22) Tel-Zrab-1, 23) Derro-2, 24) Sirom-1, 25) Sirom-4, 26) Rakka-1, 27) Kenaan-1, 28) Maskane-1, 29) Aleppo-1, 30) Azar-1, 31) Job-Hanem-1, 32) Sfaye-1, 33) Amala-1, 34) Zidan-1, 35) Habari-1, 36) Rasafa-1, 37) Didi-1, 38) Soukne-1, 39) Sharife-1, 40) Kariaten-1, 41) Tanf-1. Sections: I) Hazzro, II) Harbol, III) Iraqi Kurdistan.

Limestones. These include chemogenic, organogenic, and clay varieties.

Chemogenic limestones are formed by direct precipitation of calcium carbonate from water solution. These are subdivided into pelitomorphous, microglobular (lumpy), and micro-, small-, medium-, and coarse-grained.

Pelitomorphous limestones consist mainly of small isometric crystals. Individual crystals or crystal groups of larger dimension are found sometimes. Most of these are found, apparently, due to postsedimentary recrystallization. Rock dolomitization often evolves in this pattern with formation of small intermediate areas composed of magnesian calcite or dolomite. Kaleda [8] suggested that this recrystallization and dolomitization of limestones may be due to the uneven distribution of impurities in the general mass. Until now, there has been no consensus of the genesis of pelitomorphous limestones. Some believe that these rocks were largely formed as a result of bacterial activity [6, 7]. However, most investigators attribute pelitomorphous limestones to direct chemical precipitation [9, 10]. In our view, these limestones were probably formed by chemical precipitation in the Triassic marine basin of Syria. Among other things, this is suggested by their close paragenetic relationship with dolomites and sulfates. Pelitomorphous limestones are widespread also in the Amanus shale suite. Although in subordinate amounts, they are present practically in the entire profile. They are confined to middle portions of section members and are associated through gradual transitions with coarser-grained limestones.

Microglobular (lumpy) limestones consist of small rounded pellets (globules) of pelite-like limestone cemented by small-grained or pelitomorphous calcite (Fig. 3a, b). The genesis of these limestones is not fully clear. Khvorova [12] believes that blue-green algae, boring algae, and burrowing organisms played an important role in the formation of these limestones. In our view, globular limestones from the Syrian Triassic were formed by chemical precipitation of calcite and formation of small pellets by coagulation. This view has been supported by other investigators. Syundyukov [9] notes that the chemogenic origin of globular limestones is indirectly proved by the absence of traces of vital functions of organisms in the pellets.

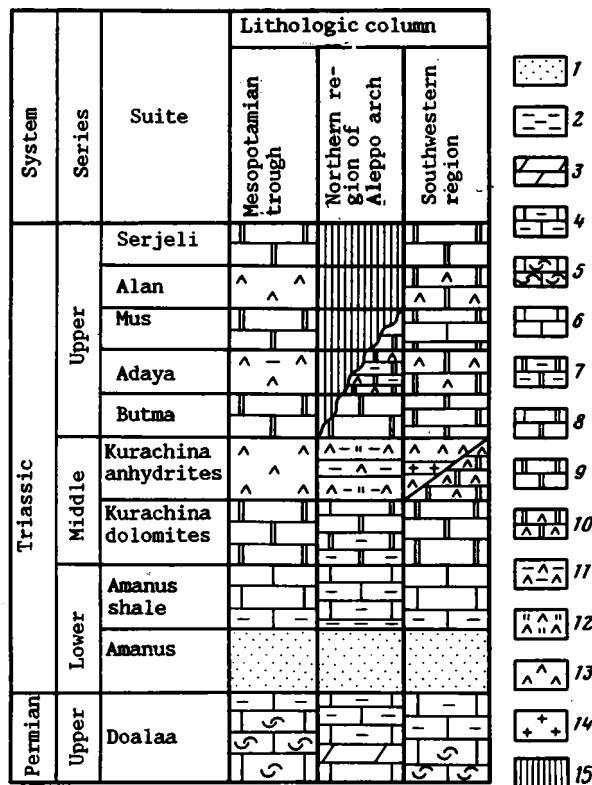


Fig. 2. Schematic separation of the Triassic in Syria. Lithologic rock types: 1) sandstone, 2) argillite, 3) marlstone, 4) clayey limestone, 5) organogenic limestone, 6) limestone, 7) clayey dolomite, 8) calcareous dolomite, 9) dolomite, 10) anhydritic dolomite, 11) clayey anhydrite, 12) dolomitic anhydrite, 13) anhydrite, 14) salt, 15) areas with no sediment accumulation.

Micro- and small-grained limestones consist mostly of microcrystalline isometric calcite. They are often dolomitized (see Fig. 3c). The top part of the Amanus shale suite consists almost entirely of these rocks. They also participate in the section of the Upper and Middle Triassic as thin members. A small concentration of impurities is typical for these rocks. The impurities are represented by rare quartz grains of silt size or clay matter distributed in thin interlayers.

Medium-grained limestones consist of isometric, less often webbed, calcite crystals. They are found through the Triassic section, often unevenly dolomitized. Along with microgranular limestones, they constitute quite thick members. As impurities, the medium-grained limestones contain single quartz fragments of silt size. These are of an angular, elongate, or isometric shape. The limestones also contain galuconite grains. There are 2-3 mm interlayers enriched in clay matter.

Coarse-grained limestones are of a limited occurrence. They are found only in the Amanus shale suite. The genesis of these limestones is attributed to recrystallization of smaller-grained varieties, indirectly supported by their tendency to be associated with heightened rock fissuring zones.

Organogenic Limestones. These include foraminiferal, organogenic-clastic and stromatolitic limestones.

Foraminiferal limestones largely consist (more than 50%) of remains of foraminiferal shells cemented by micro- and small-grained calcite. Neomorphous rhombic dolomite crystals have developed after cement in many specimens. Characteristically, the rocks contain minor amounts (single-digit percentage points) of terrigenous matter (see Fig. 3d). Foraminiferal limestones are widespread in rocks of the Doalaa suite.

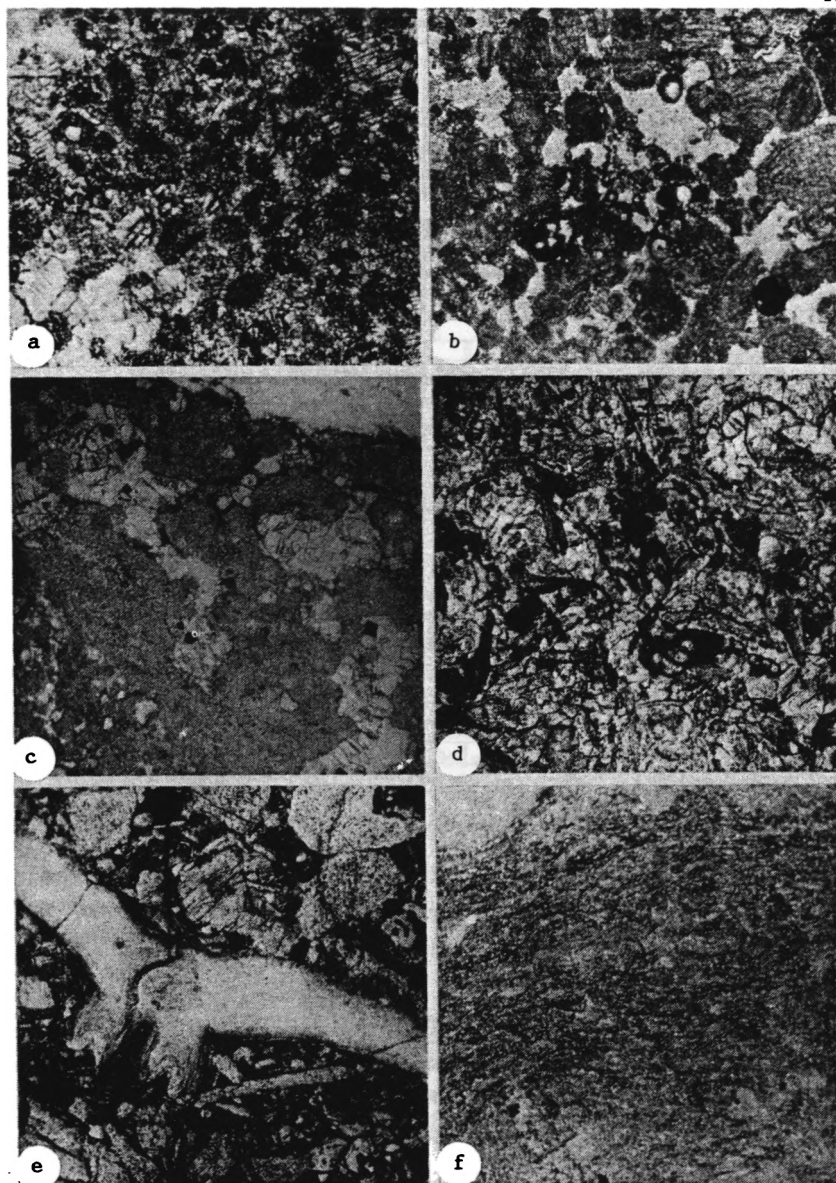


Fig. 3. Limestone: a) organogenic, microglobular (lumpy), dolomitized; lighter anhydrite neomorphs are seen in pores; b) organogenic, microlumpy (globular), fully dolomitized limestone; dolomititic cement replaced by anhydrite; c) micro- and small-grained dolomitized limestone; authigenic anhydrite developed after pores; d) foraminiferal dolomitized limestone; e) organogenic-clastic; f) stromatolitic limestone.

Organogenic-clastic limestones consist of brachiopod, echinoderm, and foraminiferal fragments, and of carbonate rocks and cement (Fig. 3e). Elongate or isometric carbonate fragments are made up of pelitomorphic or micro- and small-grained limestone. Single glauconite grains of medium or good rounding are also found. The rock cement consists of small- and medium-grained calcite. These limestones are characteristic for rocks of the Doalaa suite. Single interlayers are also found in the Amanus shale suite, where the carbonate fragment concentration rises to 40%.

Stromatolitic limestones occur as small interlayers in virtually all Lower Mesozoic suites, especially in carbonates underlying evaporitic beds. They exhibit wavy lenslike stratification underlined by alternation of light and darker interlayers, reflecting a heightened content of coaly matter (see Fig. 3f). The stromatolitic limestones are often considerably dolomitized. The dolomitization is uniform and penetrates the entire volume.

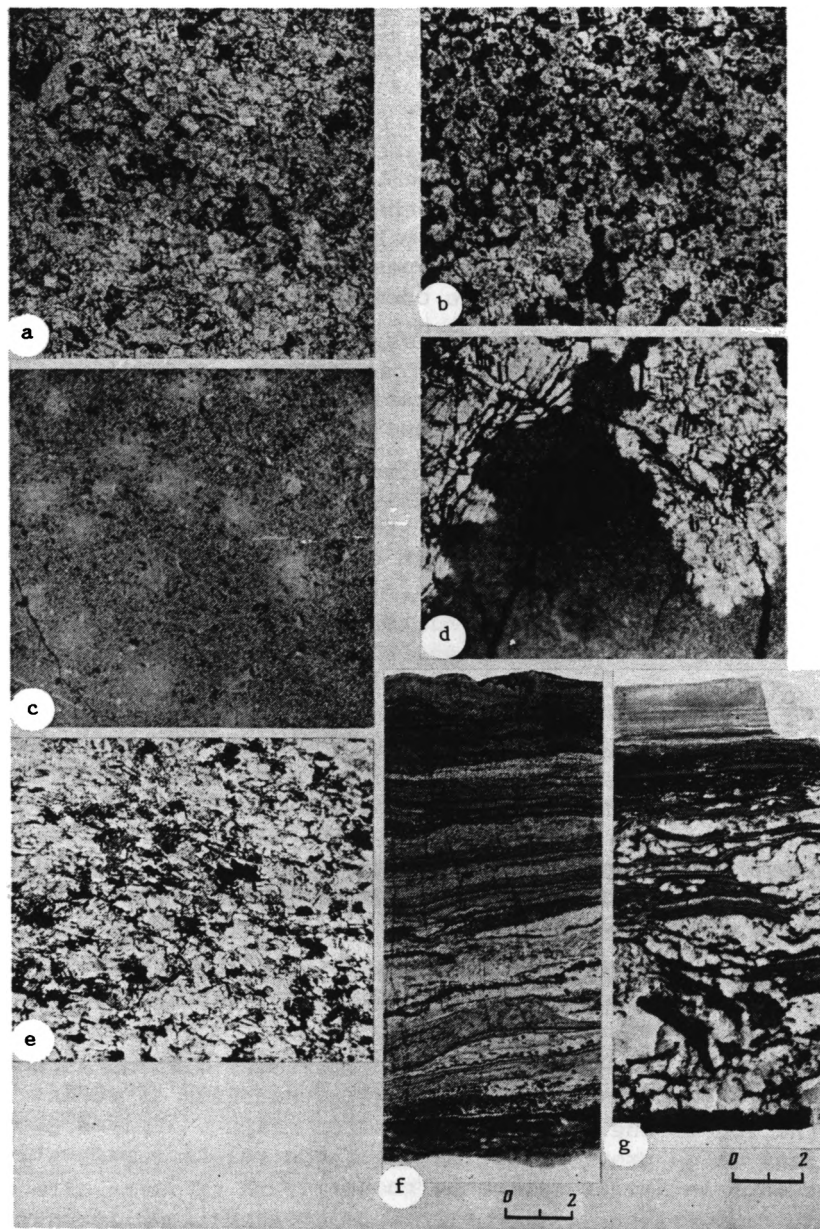


Fig. 4. Dolomite: a) organogenic, medium- and coarse-grained; organogenic remains (foraminiferal?) are virtually recrystallized and replaced by dolomite (interstitial spaces filled with bituminous matter); b) medium-grained organogenic (foraminiferal?), bituminous; c) calcareous microgranular (in a dyed polished section, light areas are represented by authigenic dolomite); d) microgranular (anhydrite neomorphs confined to cracks; e) microgranular (darker interlayers), interstratified with medium-grained irregularly structured anhydrites; f) algal (?), fine-layered with stratification underlined by the distribution of dark organic matter (upper half); g) anhydritic dolomite.

Clayey Limestones. Clayey limestones contain clay matter from 5 to 30%. These rocks are usually of a pelitomorphic texture. The rock structure is massive. With clayey dolomites they form thin (0.5-2 cm) interlayers. In Amanus shale rocks at Qasmuka, members (tens of meters thick) are found which consist of clayey limestones. In the Permian of the Doalaa suite, clayey limestones are found together with marlstones. They are connected with the latter by gradual transitions and are the predominant rock type.

Dolomites. Besides sulfates, dolomites are the main rock type. Dolomite genesis remains a largely unresolved problem, despite numerous investigations in the USSR and other countries [1-4, 10, 14, 15].

Available data do not afford an unequivocal determination whether Syrian Triassic dolomites are of a primary or secondary genesis. Describing Mesozoic sections, we continually observed limestones or magnesian limestones dolomitized to various degrees. From an analysis of data in the course of the study, we came to the view that for most dolomites the dolomitization process probably occurred at the boundary between sedimentation and diagenesis. Comparative analysis of textural and structural types of limestones and dolomites has not discovered any basic differences (aside from the chemical composition).

Dolomites are subdivided into chemogenic, organogenic, and clastic varieties, which are similar texturally to the respective limestone types. In addition, the following dolomite types have been described, where dolomitization was simultaneous with recrystallization; the secondary origin of these units is beyond question.

Medium- and coarse-grained dolomites are homogeneous, unstratified, and virtually free of admixture of other carbonates. The rock consists of medium- and coarse-grained regularly shaped crystals, with relicts of organogenic textures. Dolomite crystal shapes of this type indicate secondary recrystallization. The origin of the rocks is attributed to dolomitization of a preexisting calcareous rock or recrystallization of a different dolomite type, because usually the rocks fail to preserve any primary genetic features. These dolomites are found in all Triassic suites of Syria (Fig. 4a).

Medium-grained dolomites are homogeneous, with pores filled with bituminous matter. Typical crystals are isometric or slightly rounded. Generally, the rock is stratified. The layers are underlined by thin interlayers of organic matter. Dolomites of this kind can make up thick strata (up to tens of meters) and usually occur above argillite or limestone layers. A thorough study revealed relicts of organogenic textures (presumably foraminiferal shells). Although the secondary nature of dolomitization of these rocks cannot be questioned, it is likely to have evolved directly after sedimentation, prior to lithification of the sediment. These dolomites are widespread in the Butma and Kurachina-dolomite suites, where they are good collectors (see Fig. 4b).

Calcareous dolomite is small-grained and unstratified (see Fig. 4c). On photographs it is marked by isometric lighter areas (with indistinct boundaries) made up of microcrystalline dolomite.

Dolomitization for micro- and small-grained limestones with no stratification and massive structure presumably occurred during diagenetic reworking of rocks. These formations typically occur at the top of the Amanus shale suite. Figure 5 gives generalized data of chemical investigations which provided estimates of the relative proportions of magnesian and calcareous components in Syrian carbonate rocks.

The first diagram (Fig. 5) includes eight fields, with composition corresponding to: 1) magnesian dolomites, 2) pure dolomites, 3) calcareous dolomites, 4) limestones and magnesian limestones; 5) clayey limestones, 6) dolomitic marlstones, 7) calcareous clays, and 8) clayey dolomites.

The second diagram (Fig. 6) distinguished five fields: 1) anhydrites, 2) dolomitic anhydrites, 3) anhydritic dolomites, 4) anhydritic limestones, and 5) anhydritic dolomites.

These data help determine the composition of rocks in various suites. We see in the diagrams that the Serjeli suite consists mainly of dolomitic marlstones, whereas in the Kurachina-dolomite suite pure dolomites are more frequent. Field 8 consists of rocks of the lower part of the Butma suite. Generally, this diagram exhibits the wide variety of dolomites in the Syrian Triassic.

Anhydrites. A microscopic investigation of evaporites determined the following types of anhydrites, relatively frequent in the profiles studied.

Medium- and small-grained anhydrite, taking up from 5 to 25% of pores, of an isometric shape, and filling inner parts of shells of microorganisms. Anhydrites of this type are found both in limestones and in dolomites virtually throughout the section. Usually, they have no distinct stratigraphic association with specific carbonate suites (see Fig. 4c).

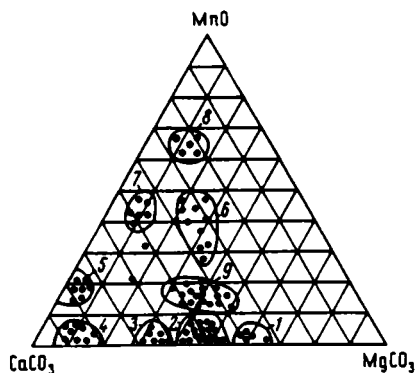


Fig. 5

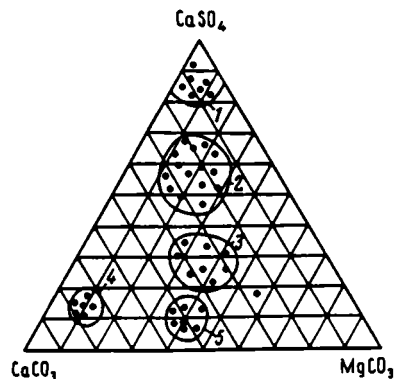


Fig. 6

Fig. 5. Composition diagram of carbonates from the Syrian Triassic: 1) magnesian dolomite, 2) pure dolomite, 3) calcareous dolomite, 4) limestone and magnesian limestone, 5) clayey limestone, 6) dolomitic marlstone, 7) marlstone, 8) calcareous clay, 9) clayey dolomite.

Fig. 6. Composition of sulfates and carbonates in the Triassic of Syria: 1) anhydrite, 2) dolomitic anhydrite, 3) anhydritic dolomite, 4) anhydritic limestone, 5) anhydritic dolomite.

The shape and texture of these anhydrites clearly determines their secondary origin. It is confirmed by secondary redistribution of sulfate matter in section, which migrated through cracks and microcracks and filling free interstitial spaces. This could have been accompanied by partial dissolution of primary carbonate matter in pores and its migration to other parts of the section.

Medium- and small-grained anhydrite (less often, coarse-grained) makes up from a few percentage points to up to 30% of the rock, filling cracks. Typical crystals are acicular; they are parallel to each other and perpendicular to the sides of the cracks.

Anhydrites of this type usually occur in dolomites overlying beds of pure anhydrites in virtually all suites – from Kurachina dolomites at the section bottom to the Serjeli suite at its top. The shape and texture of these entities clearly indicate their origin. Most likely, they were formed during transition of sedimentary gypsum to anhydrite, when released water could easily migrate up the section, carrying some sulfate matter and depositing it along the path in carbonate rock cracks (see Fig. 4d).

Anhydrite with a disordered structure, medium-grained, making up from 50 to 80% of rock, often forms portions with starlike distribution of crystals similar to the first type. The rock has preserved areas of small-grained carbonate heavily corroded on edges. Such anhydrites are usually found in the direct vicinity and above thick anhydrite beds or in dolomite layers situated inside anhydritic strata. They are especially common in the Kurachina-anhydrite, Butma, Aday, Alan, and Serjeli suites. Their secondary origin is indicated by preserved relict portions of carbonate, a largely disorderly texture with frequent starlike areas, the absence of stratification, the location in the section, and usually, thin layers. It is also possible that this rock type was formed by postsedimentary migration and redistribution of sulfate matter in the section (see Fig. 4e).

Medium- and small-grained anhydrite exhibits ordered mutually parallel (to rock stratification) orientation of crystals. The rock is typically gray, horizontally wavy-layered, sometimes with argillite interlayers. Such anhydrites form thick beds, which make up most of the Kurachina-anhydrite, Adan, and Alan suites. Similar but thinner beds are also found in the Kurachina-dolomite, Butma, and Serjeli suites. The primary genesis of these formations is suggested by their stratification, the orientation of crystals, the thickness of the beds they form, and have widespread blanket occurrence on almost the entire territory of Syria (see Fig. 4g).

These rocks were apparently formed in a shallow-sea basin during limited inflow of normally saline water and an inflow of water from inland. Typically, they are associated with the basin parts that are farthest from the shore.

Medium- and small-grained anhydrite, with ordered texture, enriched in clay matter, with thin interlayers of anhydritic argillites and argillites with imprints of large (1-2 mm) gypsum and anhydrite crystals, contains crystals of a mutually parallel orientation along the rock stratification. The rocks are mostly of gray, dark-gray, and greenish color; they are striated and horizontally stratified. Such anhydrites usually are not thick (less than 10 m). They occur directly on stromatolitic dolomitized limestones. This fact suggests that these anhydrites were formed near the coast in lagoons and littoral shallow-sea conditions, probably with periodic drying (see Fig. 4f).

Sandstones. These rocks constitute the Amanus suite and partly the Amanus shale suite. A stable mainly quartzose composition is a typical feature of these sandstones.

In the northeastern Syria the sandstones are small- and medium-grained, of a gray color, unstratified, and of medium sorting. They consist of poorly rounded or unrounded isometric or triangular quartz fragments. In the section of the Salhiya-1 well, semirounded and tabellar fragments of feldspars represented by potash feldspar and albite have also been found. Their concentration does not exceed 10%. Sandstone cement varies along the profile. At the bottom of the Amanus suite it is calcareous and of basaltic type; up the section it gradually gives way to clayey-calcareous and clayey cement of interstitial type. Iron hydroxides also appear in the cement; the rocks exhibit a brownish pigmentation and conformal texture. In the Amanus shale suite small-grained quartzose sandstone with poor rounding and medium fragment sorting constitutes thick (up to 1 m) layers interstratified with siltstone and argillites.

Argillites. Clay rocks that occupy a subordinate position in the Triassic section are represented by dark-gray and variegated argillites of the Amanus shale suite and dark-gray and black dolomitic argillites found in thin interlayers in the Kurachina-anhydrite, Butma, Aday, Mus, Alan, and Serjeli suites. The argillites of the Amanus shale suite are thick (tens of meters) with a mixed composition due to the presence of psammitic or carbonate material; they also contain frequent interlayers of siltstones, sandstones, or limestones. They are diversely colored by iron hydroxides, in turn, indicating littoral or continental formation conditions.

Carbonate and sulfate beds contain clay rocks as thin (1-2 cm) interlayers mostly represented by dark dolomitic argillites. Carbonate content in them is as high as 40%. According to x-ray studies, clay minerals include kaolinite, hydromica, and mixed layer hydromica-smectite formations. The latter are prevalent at the section bottom in the Amanus shale suite; kaolinite begins to have a greater role up to the section.

Lithologic and Facies Characteristics of the Triassic

Lithologic, petrographic, and genetic studies of rocks suggest distinguishing two groups: continental and marine.

Continental sediments occur in limited locations. They are mainly confined to perimeters of large elevations, such as Rutba, Iraq, Kamyshli-Sirom, Aleppo, and others, which have affected greatly the course of sedimentation, especially during the Early Triassic (Fig. 7). The margins of these uplands were the location of microfacies (landscape zone) of marine valleys. They include the facies of sand-clay sediments of littoral accumulation planes, represented by variegated quartzose medium- and small-grained sandstones with poor sorting and medium rounding, including variegated argillite and siltstone interlayers. These sandstones are widespread in almost all lower Triassic sections in layers of 20 to 30 m or more.

Marine sediments represent three macrofacies, according to their formation environments. Sediments of the first macrofacies were formed in an open littoral shallow sea with relatively intense water dynamics. Macrofacies of semiinsulated coastal shallow-water areas are characterized by stagnant or partially stagnant marine shoals. Finally, at a considerable distance from the shore, with the sharp reduction of influx of terrigenous sand-silt material, clay, carbonate, sulfatic, and halitic sediments of the macrofacies of distant areas of a shallow marine basin were formed.

Macrofacies of sediments of open littoral shallow-sea basins is associated genetically with terrigenous rocks of littoral accumulation planes. Coastward, they are gradually succeeded by these terrigenous rocks. They are distinguished from continental complexes by a gray or light-gray pigmentation, less coarse granulometry, and the presence of carbonate

material. These sediments cover a large area, contouring island rises. They are widespread at some distance offshore.

The macrofacies is subdivided into two facies. The facies of a mobile littoral shallow-sea of an epicontinental marine basin is represented by quartzose sandstones and silty sediments with an admixture of carbonate matter. It consists mainly of sand of medium and good sorting, with a light-gray pigmentation, occurring in horizontal or horizontal-wavy layers, with coalified plant remains. Laterally, it makes transition to continental facies on one side and remote marine basin facies on the other. This facies mainly occurs at the bottom of the Lower Triassic section. The facies of low mobility littoral shoals of the epicontinental marine basin is represented mainly by clays, marlstones, and microgranular dense limestone with some silt. They were formed in the transitional zone between coastal mobile shoals and remote offshore areas. The sediments are of a clay-carbonate composition, with dark-gray pigmentation and horizontal or horizontal-wavy stratification, containing remains and imprints of vegetation and fauna. These facies are formed in regressive parts of the Syrian Triassic section.

Macrofacies of sediments of semiisolated lagoon littoral shoals of the marine basin is typical for stagnant isolated bays and lagoons with periodically drying, littoral and shallow-sea sediments. The macrofacies is subdivided into two facies. The facies of semi-isolated lagoon and littoral marine shoals is represented by magnesian carbonate sediments: micro- and small-grained dolomites, calcareous dolomites, and clay dolomites formed in littoral shallow-sea aseminsulated parts of the marine basin that cut into the land, with limited seawater influx in arid climates. They consist mainly of dolomites with gray or dark-gray pigmentation and contain algal stromatolitic formations, with finely striated structure or a "bird's eye" structure. The sediments occur in thin horizontal layers at the bottom of a lagoon-sebkha complex. The formations are widespread in the Middle and Upper Triassic.

The facies of periodically isolated lagoons and sebkhas is represented by dark-gray and black argillites with gypsum crystal imprints and gray, dark-gray and lumpy anhydrites of an unordered structure formed in stagnant and partly or completely isolated sea areas in lagoons and sebkhas with periodic influx of seawater in an arid climate. Vast inundated shallow plane areas were apparently formed along the coast during regression periods. Intense evaporitization and precipitation of sulfate formations evolved in these regions. Sediments of this facies are characterized by a clay-sulfate composition, globular or disordered sulfate structure, fine horizontal stratification of argillites, an absence of organic remains, dark or spotty pigmentation, and occurrence at the tops of lagoon-sebkha complexes. Rocks of this facies are widespread in the Middle and Upper Triassic over the Syrian territory.

Macrofacies of sediments of offshore shallow parts of the marine basin refers to the most distant regions of the basin, where calm hydrodynamic conditions of the sea were predominant. The sediments of this macrofacies differ from the preceding varieties by a biogenic genesis and small amounts of terrigenous matter. They occupy vast areas, both in central Syria and in the country's northeast. The macrofacies is subdivided into the facies of open sea offshore areas represented by calcareous argillite, gray and dark-gray small- or microgranular clay limestones, with rare foraminiferal remains, and single quartz and galuconite fragments occurring in horizontal and wavy stratified layers. The sediments of this facies were formed in the remote parts of the sea basin in calm hydrodynamic and normal hydrochemical conditions. They are characterized by a calcareous clay composition, finely dispersed material, some faunistic remains, an absence of detritus or clastics, gray and dark-gray color, thin horizontal or horizontal-wavy stratification, gradual transition to other marine facies, wide area of occurrence, and presence in the top parts of transgressive cycles. These sediments are virtually ubiquitous in the upper parts of the Lower Triassic. The second facies is the facies of sediments of an offshore partly isolated sea basin, represented by dolomites, calcareous dolomites, sometimes anhydritic dolomites, with diverse textures and structures - from pelitomorphic to coarse-crystalline and from massive to striated, variegated, and globular. These rocks were formed far from the shore in an arid climate, with limited normal seawater influx and no supply of clastics. They are widespread on the entire territory of Syria and can be found in all the suites of the Middle and Upper Triassic. The rocks are of a predominantly dolomitic composition with a slight admixture of clay material. They are light-gray and gray and occur in horizontal layers. The rocks contain remains of poorly preserved planktonic organisms on extensive areas and paragenetic correlation with sediments of other marine facies. The facies is found at the bottom of regressive cycles.

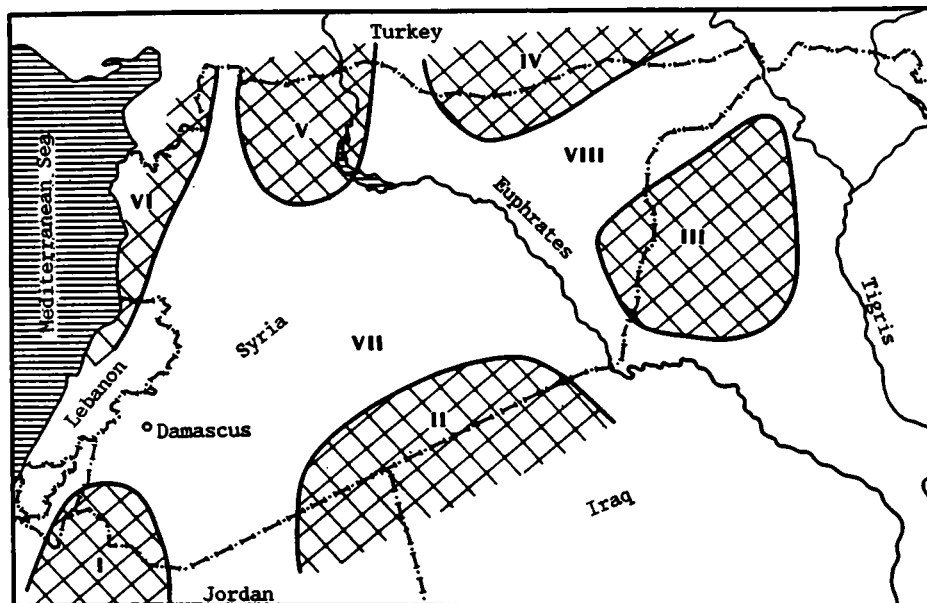


Fig. 7. Paleomorphostructural map of the Triassic. Uplands: I) Jordan, II) Rutba, III) Iraq, IV) Kmyshla, V) Aleppo, VI) Latak-Kil. Troughs: VII) Central Syrian, VIII) Mesopotamian (northeastern).

The facies of sediments from offshore isolated marine basins are represented by gray anhydrites, which are small- and medium-grained rocks of a striated or lenticular structure; it also contains medium- and coarse-grained halites. The facies was formed in central parts of the sea basin during periods of intense evaporitization under no or restricted influx of seawater. The rocks are of a typical sulfatic-magnesian or halitic composition, with virtually no terrigenous material. The layers reach tens of meters in thickness, contain no organic remains, and occur at tops of regressive cycles. The facies is distinguished from littoral evaporites by its greater thickness and an absence of clay material. In plan, it forms a concentric pattern with halitic and sulfatic sediments at the center and magnesian complexes on the perimeter. The formations of this facies make up the Kurachina-anhydrite, Alan, and Adaya suites.

From an analysis of the genetic characteristics of the Triassic sediments, their formation settings were reconstructed. The Syrian Triassic is associated with four macrofacies, each corresponding to a particular landscape zone with specific sedimentation environments. The genetic features were reflected primarily in the composition of sediments, lithologic rock types, structural and textural features, and other characteristics.

The lithology and genetic properties of the Triassic indicate a regular distribution of various sediment types in the basin and a directed evolution of sedimentation during the Triassic in Syria's territory.

The terrigenous rock complexes of littoral and near-sea facies were formed in the direct vicinity of the coastline and were widespread during the Early Triassic, when a new stage in the region's geologic history began after an interruption in sedimentation. By the Middle Triassic, most of the territory was covered by a shallow sea, and carbonate sediments of offshore marine facies were formed virtually everywhere. The sole exception was paleouplands, such as Aleppo and Rutba, where specific conditions were created and sulfate-clay sediments of the lagoon-sebkha facies were formed. During local regressions in the Middle and Upper Triassic, stagnant evaporitic settings developed in the entire Syrian basin, bounded by a ring of island uplands; at that time, magnesian carbonates and sulfates were deposited all over the territory.

Generally, the evolution of the sedimentation basin in Syria followed the sequence from major transgression in the Lower Triassic with development of facies diversity to a gradual contraction of the basin in the Middle and Upper Triassic and accumulation of regressive sediments.

LITERATURE CITED

1. S. G. Vishnyakov, "Genetic types of dolomitic rocks in the northwestern margin of the Russian platform," in: Types of Dolomite Rocks and Their Genesis [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1956), pp. 78-86.
2. S. G. Vishnyakov, "Classification of sedimentary rocks," in: Proceedings of an International Conference on the Geology of Fossil Minerals from Chernozem Regions [in Russian], Izd. Voronezh. Un-ta, Voronezh (1957), pp. 25-34.
3. M. N. Zharkov, "Evolution of salt accumulation in geologic history," in: Problems of General and Regional Geology [in Russian], Nauka, Novosibirsk (1971), pp. 260-299.
4. M. N. Zharkov, History of Paleozoic Salt Accumulation [in Russian], Nauka, Novosibirsk (1978).
5. Yu. A. Zhemchuzhnikov, V. S. Yablokov, L. I. Bogolyubova, et al., "Structure and accumulation settings of main coal seams of the Middle Carboniferous in the Donets Basin," Tr. GIN Akad. Nauk SSSR, Issue 15, Vol. 4, No. 1 (1959).
6. B. L. Il'chenko, "Biogenic formation of calcium carbonate," Mikrobiologiya, 17, No. 2, 8-14 (1984).
7. V. O. Kalinenko, "Bacterial precipitation of calcium in the sea," Tr. Okeanol. Inst. Akad. Nauk SSSR, No. 3, 42-47 (1949).
8. G. A. Kaleda, "Recrystallization of carbonate rocks," Voprosy Mineralogii Osadochnykh Obrazovaniy, Issue 2, 87-91 (1955).
9. A. Z. Syundyukov, Lithology, Facies, and Oil and Gas Prospects of Carbonate Rocks in Western Bashkiria [in Russian], Nauka, Moscow (1975).
10. N. M. Strakhov, Fundamentals of the Theory of Lithogenesis, Vol. 3 [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1962).
11. P. P. Timofeev, "Jurassic coal-bearing formations of South Siberia and its development environments," Tr. GIN Akad. Nauk SSSR, Issue 198 (1969).
12. I. V. Khvorova, Atlas of Carbonate Rocks of the Middle and Upper Carboniferous of the Russian Platform [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1958).
13. K. E. Chave, "Aspects of the biochemistry of calcareous sediments and rocks," J. Geol., 62, No. 6, 587-599 (1954).
14. P. R. Siegel, "High pH and primary dolomites," Sedimentation, 5, No. 3, 255-261 (1965).
15. J. I. Wilson, "Characteristics of carbonate platform margins," Bull. Am. Assoc. Petrol. Geol., 58, 137-146 (1974).

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